SIMULATION OF GROUND-WATER FLOW OF THE COASTAL PLAIN AQUIFERS IN PARTS OF MARYLAND, DELAWARE, AND THE DISTRICT OF COLUMBIA

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1404–J

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Simulation of Ground-Water Flow of the Coastal Plain Aquifers in Parts of Maryland, Delaware, and the District of Columbia

By WILLIAM B. FLECK *and* DON A. VROBLESKY

REGIONAL AQUIFER-SYSTEM ANALYSIS—
NORTHERN ATLANTIC COASTAL PLAIN

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

The Regional Aquifer-System Analysis (RASA) Program was started in 1978 following a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States. The RASA Program represents a systematic effort to study a number of the Nation's most important aquifer systems, which in aggregate underlie much of the country and which represent an important component of the Nation's total water supply. In general, the boundaries of these studies are identified by the hydrologic extent of each system and accordingly transcend the political subdivisions to which investigations have often arbitrarily been limited in the past. The broad objective for each study is to assemble geologic, hydrologic, and geochemical information, to analyze and develop an understanding of the system, and to develop predictive capabilities that will contribute to the effective management of the system. The use of computer simulation is an important element of the RASA studies, both to develop an understanding of the natural, undisturbed hydrologic system and the changes brought about in it by human activities, and to provide a means of predicting the regional effects of future pumping or other stresses.

The final interpretive results of the RASA Program are presented in a series of U.S. Geological Survey Professional Papers that describe the geology, hydrology, and geochemistry of each regional aquifer system. Each study within the RASA Program is assigned a single Professional Paper number, and where the volume of interpretive material warrants, separate topical chapters that consider the principal elements of the investigation may be published. The series of RASA interpretive reports begins with Professional Paper 1400 and thereafter will continue in numerical sequence as the interpretive products of subsequent studies become available.

Gordon P. Eaton
Director

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(Plates are in pocket)

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CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATIONS

For the convenience of readers who may prefer to use metric (International System) units rather than the inch-pound units used in this report, values may be converted by using the following factors:

Multiply inch-pound unit	By	To obtain metric unit
inch (in.)	25.4	millimeter (mm)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
foot (ft)	0.3048	meter (m)
foot per year (ft/yr)	0.3048	meter per year (m/yr)
foot per day (ft/d)	0.3048	meter per day (m/d)
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
cubic foot per second per cubic foot [$(ft^3/s)/ft^3$]	1	cubic meter per second per cubic meter [$(m^3/s)/m^3$]
cubic foot per second per square mile [$(ft^3/s)/mi^2$]	0.01093	cubic meter per second per square kilometer [$(m^3/s)/km^2$]
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.509	square kilometer (km^2)

Chemical concentration is given in metric units and is expressed in milligrams per liter (mg/L).

The volumetric flux unit used in this report is cubic feet per second (ft^3/s). To convert this unit to gallons per day (gal/d), multiply by 646,323; to million gallons per day (Mgal/d), by 0.646.

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929), a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called *Sea Level Datum of 1929*.

SIMULATION OF GROUND-WATER FLOW OF THE COASTAL PLAIN AQUIFERS IN PARTS OF MARYLAND, DELAWARE, AND THE DISTRICT OF COLUMBIA

By WILLIAM B. FLECK and DON A. VROBLESKY

ABSTRACT

The Coastal Plain sediments of Maryland, Delaware, and the District of Columbia consist of a thick, complex sequence of sand, gravel, silt, and clay, which range in age from Cretaceous to Quaternary. These sediments generally crop out in broad bands parallel to the Fall Line and dip at a low angle southeastward.

These sediments were divided into 10 regional aquifers and 9 intervening confining units. The aquifers, in ascending order, are the Patuxent, Patapsco, Magothy, Matawan, Severn, Aquia-Rancocas, Piney Point–Nanjemoy, lower Chesapeake, upper Chesapeake, and surficial. The confining units are the Potomac, Patapsco, Matawan, Severn, lower Brightseat, Nanjemoy-Marlboro, lower Chesapeake, St. Marys, and upper Chesapeake.

A quasi-three-dimensional model for ground-water flow was constructed to simulate the flow system of the 10 aquifers and 9 confining units. The model used a rectangular grid of 42 rows by 36 columns of square cells, each cell being 3.5 miles on a side. No-flow boundary conditions were assumed at (1) the top of the underlying crystalline rocks, (2) the featheredge of the outcrop areas, and (3) the downdip limit of the freshwater flow boundary (line where chloride concentration is 10,000 milligrams per liter). Specified flux, as provided by a model of the entire northern Atlantic Coastal Plain, was used along the northern boundary in New Jersey and the southern boundary in Virginia. For the simulation of prepumping conditions, the water table was a specified-head boundary. For the simulations of transient conditions for the period 1900–80, a set of surface-water cells was added to create a specified-head boundary condition. The water table was then allowed to fluctuate in response to stresses on the aquifer system. Model input included transmissivities, storage coefficients, leakances, areal ground-water recharge, and ground-water withdrawals.

Maximum transmissivity values used in the model were 23,200, 19,700, and 13,900 feet squared per day in the Patapsco, surficial, and Patuxent aquifers, respectively. Leakance values ranged from 10^{-14} cubic feet per second per cubic foot [$(ft^3/s)/ft^3$] in the confining units overlying the Potomac, Nanjemoy-Marlboro, and lower Chesapeake aquifers, to 10^{-3} $(ft^3/s)/ft^3$ in other confining units. Total simulated pumpage for 1978–80 was 204 cubic feet per second (ft^3/s), which represents about 60 percent of the actual pumpage. This simulated pumpage included all major public supplies and industrial users. Minor irrigation, rural, and domestic supplies were not simulated because they are generally from the surficial aquifer, are areally distributed, and have locally affected water levels in the surficial aquifer. The largest pumpage withdrawals were 55, 51, and 39 ft^3/s in the Patuxent, Patapsco, and surficial aquifers, respectively. As a result of this pumpage, large regional cones of depression have developed. The largest cones occur in the Piney Point, lower Chesapeake, Patuxent, and Magothy aquifers, whose drawdowns are about 150, 100, 130, and 80 feet, respectively.

The model was calibrated for both prepumping conditions (1900) and transient pumping conditions (1900–80). Calibration was evaluated for each aquifer by comparing simulated heads with measured heads using 121 hydrographs and by comparing simulated heads with potentiometric-surface maps for both prepumping and transient conditions. The criterion used to evaluate the goodness of fit with measured heads was that simulated heads should be within 5 percent of the total head range for the whole flow system (18 feet).

The simulated hydrologic budget for prepumping conditions indicates that the principal source (99.8 percent) of ground water is recharge from precipitation, which was 11,419 ft^3/s. The largest quantity of recharge for any aquifer was 8,384 ft^3/s, which was applied to the surficial aquifer. The principal sink (99.6 percent) for the model was discharge to surface-water bodies and equaled 11,392 ft^3/s. Average ground-water flow velocities ranged from 0.01 feet per day in the Matawan and Severn aquifers to 0.19 feet per day in the surficial aquifer. The maximum ground-water flow velocity for any given cell was 5.3 feet per day in the Patapsco aquifer.

An analysis of the model for the 1978–80 pumping period of the transient simulation indicated that the major source of pumpage (about 204 ft^3/s) was a reduction in discharge to surface-water bodies of 184 ft^3/s from the simulation of prepumping conditions. An analysis of ground-water flow directions indicated major shifts in regional flow directions in the Patuxent, Magothy, Piney Point–Nanjemoy, and Aquia-Rancocas aquifers. Changes in ground-water flow velocities occurred in several aquifers. For example, in the Patapsco aquifer, velocities increased threefold in northern Delaware, and in the Piney Point–Nanjemoy aquifer the average velocity increased about 50 percent. Flow through confining units into the Piney Point–Nanjemoy and lower Chesapeake aquifers increased by 9 ft^3/s, and for the upper Chesapeake aquifer by 13 ft^3/s. For the surficial aquifer, flow out and through the underlying confining unit increased by 40 ft^3/s.

A sensitivity analysis of the model indicated that a 20-percent increase in pumpage resulted in heads being decreased by as much as 15 feet. A decrease of model transmissivities by a factor of 0.2 in a cell located near pumped wells changed the altitude of the head from 40 to 205 feet below sea level.

INTRODUCTION

BACKGROUND

The northern Atlantic Coastal Plain is both densely populated and heavily industrialized. As the limit of available surface-water supplies is approached, an increasing demand for ground water, in turn, will necessitate effective ground-water management. For example, use of ground water within the project area increased about 60 percent from 1970 to 1980, resulting in large water-level declines, especially in northern Delaware and southern Maryland. An effective tool in the management of ground-water resources is the digital model that can both simulate the flow system and be used to predict water-level changes for various pumpage conditions. In the past, a number of areally small models were developed to simulate one or several aquifers to evaluate future pumping effects. Most of these models have not adequately addressed the influences of a larger regional flow system on the modeled areas.

The Regional Aquifer-System Analysis (RASA) study of the northern Atlantic Coastal Plain, initiated during fiscal year 1979, includes the area from Long Island, N.Y., to the North Carolina–South Carolina border (fig. 1). This study area was further divided into five geographical subregions. Each of these separate studies developed a model that simulated the flow system within the Coastal Plain sediments for each subregion, the present one being a model for the subregion encompassing parts of Maryland, Delaware, and the District of Columbia (fig. 1). A coordinating study developed a regional model of the entire northern Atlantic Coastal Plain (Leahy and Martin, 1994).

PURPOSE AND SCOPE

The purpose of this report is to describe regional ground-water flow in a complex, interrelated series of aquifers and confining units that comprise the Coastal Plain sediments of Maryland, Delaware, and the District of Columbia. Available hydrogeologic data were incorporated into a 10-layer digital model of ground-water flow that was calibrated under prepumping and transient-pumping conditions. The model was then used to determine the direction and magnitude of ground-water flow under both prepumping and 1980 pumpage conditions.

Data collected from existing sources were used to describe the geometry of the hydrogeologic system (Vroblesky and Fleck, 1991) and to determine the flow-system boundaries and hydrologic properties used in the model. Records of historical pumpage were collected both from existing data files and from individual users (Wheeler and Wilde, 1989).

A quasi-three-dimensional digital-model program developed by the U.S. Geological Survey (Trescott, 1975) was used to simulate flow within the ground-water system of the study area. This program simulates horizontal flow within aquifers and vertical flow through intervening confining units.

PREVIOUS MODELING INVESTIGATIONS

Previous modeling in the study area provided useful data for model input as well as potentiometric-surface maps that aided in the calibration of this model. These previous studies included investigations of the Piney Point and Cheswold aquifers in Delaware (Leahy, 1979, 1982a), the Potomac Formation in Delaware (Martin, 1984), the Miocene aquifers of southern Delaware (Hodges, 1984), the Cretaceous aquifers of southern Maryland (U.S. Army Corps of Engineers, 1983), the Piney Point aquifer in Maryland (Williams, 1979), the Aquia-Nanjemoy system in Maryland (Chapelle and Drummond, 1983), the Magothy aquifer in southern Maryland (Mack and Mandle, 1977), the Ocean City aquifer system in Maryland (Achmad and Weigle, 1979; Weigle and Achmad, 1982), and the unconfined aquifer of central and southern Delaware (Johnston, 1977).

LOCATION AND PHYSICAL SETTING OF STUDY AREA

The study area encompasses all of the Coastal Plain of Maryland, Delaware, and the District of Columbia (fig. 1). The Coastal Plain is bounded on the west by the Fall Line, which delineates the pinch-out of unconsolidated sediments of Cretaceous age on the older consolidated rocks of the Piedmont Province. On the east, the study area is generally bounded by the Atlantic Ocean, and on the north and south, by the States of New Jersey and Virginia, respectively. The study area encompasses about 8,500 mi^2 (square miles), but for modeling purposes the digital model was extended in all directions and covered about 12,700 mi^2.

Two major bay features are the Delaware and Chesapeake Bays. The Chesapeake Bay effectively divides the area into two distinct physiographic areas. On the eastern side of the bay are the Eastern Shore of Maryland and the State of Delaware, a low, flat-lying, principally rural area. On the basis of records from four weather stations for 1951–80, annual precipitation averages about 44.1 in. (inches) (National Oceanic and Atmospheric Administration, 1981). The western side of the Chesapeake Bay is hillier and more dissected; considerably more urbanized, including the industrial and commercial centers of Baltimore, Md., and Washington, D.C.; and had an annual precipitation averaging 41.7 in/yr (inches

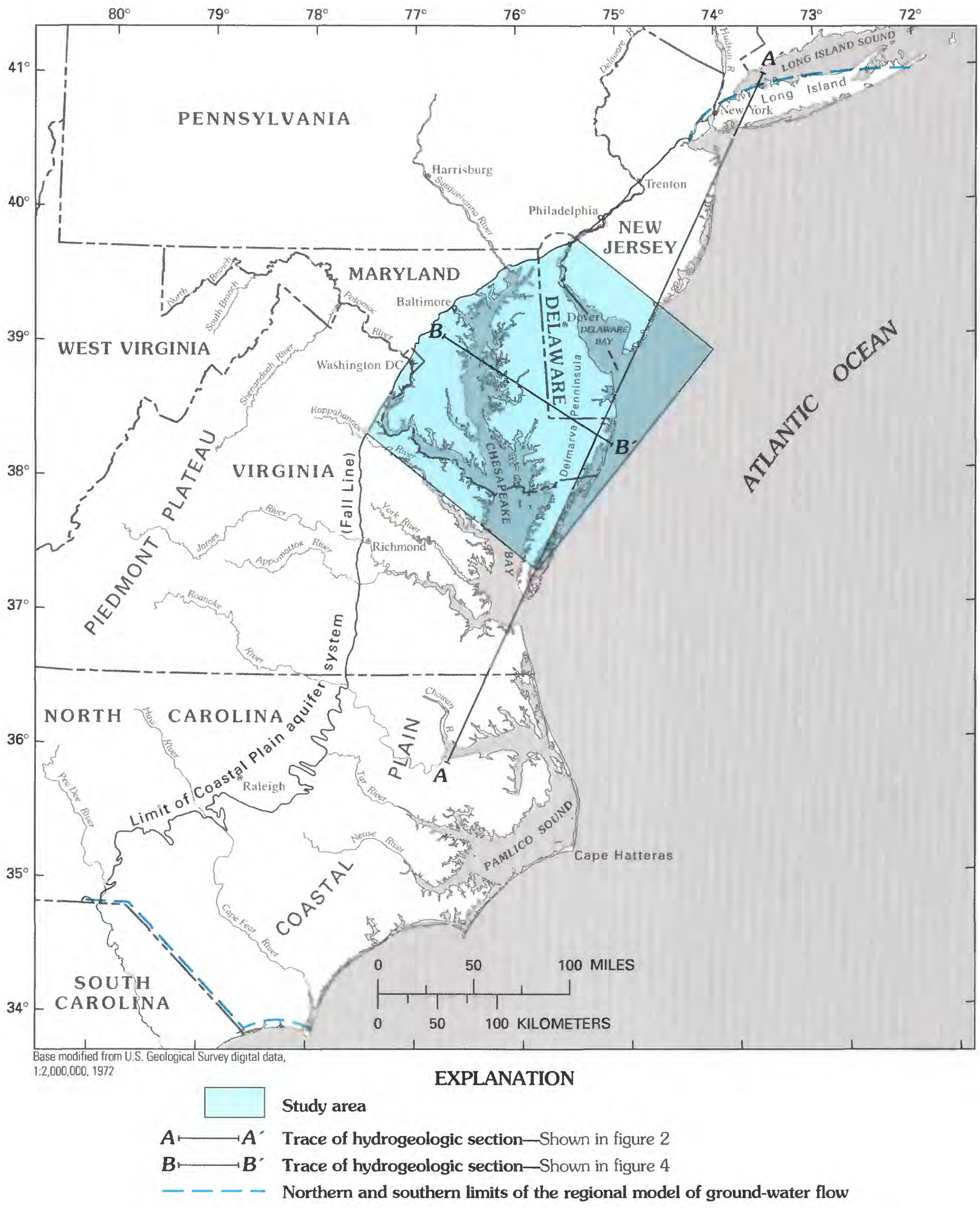


FIGURE 1.—Location of study area in the northern Atlantic Coastal Plain (modified after Leahy, 1982b, fig. 1).

per year) over 1951–80 (National Oceanic and Atmospheric Administration, 1981).

GEOLOGIC SETTING

The Coastal Plain sediments that crop out in Maryland, Delaware, and the District of Columbia range in age from Early Cretaceous to Quaternary (table 1). Only sediments of Early Cretaceous age crop out in the District of Columbia; therefore, that area is not included in table 1. With local exceptions of ferruginous or calcareous cementation, these sediments consist of unconsolidated sand, silt, clay, and gravel. Bedding strikes northeast-southwest and dips at low angles of generally less than 1° eastward to southeastward. The western boundary is defined by the Fall Line, along which the Lower Cretaceous sediments generally overlap onto the Precambrian crystalline complex of the Piedmont Province.

The crystalline rocks comprise the basement upon which the Coastal Plain sediments rest. The basement surface forms an east-plunging trough (fig. 2) known as the Salisbury Embayment. The centrally located Maryland-Delaware portion of this embayment consists principally of Lower Cretaceous sediments (about 75 percent). The Coastal Plain sediments range in thickness from a featheredge along the Fall Line to about 8,500 ft (feet) along the Maryland Atlantic Coast. As indicated in figure 2, the thickness of sediments diminishes southward through Virginia and North Carolina, and northeastward through New Jersey into New York. Figure 3 is a structure-contour map that delineates the altitude of the top of the basement within the study area.

The oldest and deepest sedimentary rocks are known from two oil test wells on the Eastern Shore of Maryland. These sediments, possibly Jurassic in age (Hansen, 1984, p. 3), are about 550 ft thick and consist of indurated basal conglomerate, sandstone, and shale intercalated with sand and shale. These sediments pinch out against the Precambrian basement complex at about 5,000 ft below sea level. Thus, the areal extent of the Jurassic is limited to an area along the Atlantic Coast of Maryland and southern Delaware.

NUMBERING SYSTEMS FOR WELLS

The wells described in this report are numbered according to a coordinate system in which the counties of Maryland and Delaware are divided into 5′ quadrangles of latitude and longitude. These numbering systems were originated by the respective State Geological Surveys. In Maryland, the first two letters of the well number identify the county in which the well is located; for example, "AA" indicates Anne Arundel County. Then a quadrangle code identifies a 5′ segment of latitude and a 5′ segment of longitude. The final digits in a Maryland code are assigned chronologically. Thus, well AA Fe 47 is the 47th well inventoried in quadrangle Fe in Anne Arundel County.

The first and second letters in a Delaware well number refer to a 5′ segment of latitude and a 5′ segment of longitude, respectively. Each 5′ quadrangle is further divided into 1′ segments. The row and column numbers of the 1′ quadrangles are the third and fourth digits, respectively. The final digits are assigned chronologically. Thus, well Gd 34-2 is the second well inventoried in row 3, column 4, of 5′ quadrangle Gd in Delaware.

DESCRIPTION OF MODEL LAYERS

The hydrogeologic framework described by the authors in the companion Professional Paper 1404–E (Vroblesky and Fleck, 1991) is the basis of the geometry for the digital flow model. The model for the northern Atlantic Coastal Plain RASA study consists of 10 layers, each of which consists of an aquifer and overlying confining unit. These 10 layers and the equivalent aquifers, confining units, and geologic formations are indicated in table 1. The areal extent of each aquifer is given on plates 1*A*, 2*A*, 3*A*, 4*A*, 4*G*, 5*A*, 6*A*, 7*A*, 8*A*, and 9*A*. Layer 10 represents the surficial aquifer and, as such, has no overlying confining unit. A description of the aquifers and confining units and a series of thickness maps for each unit are presented in Professional Paper 1404–E (Vroblesky and Fleck, 1991).

The Coastal Plain sediments of the study area were deposited in a series of changing sedimentary environments (Hansen, 1972, p. 4). Lower Cretaceous and Quaternary sediments were deposited in fluvial environments (table 1, fig. 4). The Magothy, Marlboro, and Yorktown Formations were deposited in fluviomarine environments, and the rest of the sediments were deposited in marine sedimentary environments.

LAYER 1

PATUXENT AQUIFER

Layer 1 (pl. 1*A*) represents the Patuxent Formation of Early Cretaceous age, and is referred to in this report as the "Patuxent aquifer" and the overlying "Potomac confining unit." The aquifer is bounded on the west by the Fall Line, which is the edge of the Piedmont Province. To the north and south, this layer terminates at the New Jersey and Virginia State lines. The boundary to the east represents an assumed no-flow boundary at the saltwater-

TABLE 1.—*Relation of stratigraphic units, hydrogeologic units, and equivalent layers in the computer model of the Coastal Plain of Maryland and Delaware (Vroblesky and Fleck, 1991, plate 1)*

System	Series	Stratigraphic units[1]: Maryland	Stratigraphic units[1]: Delaware	Lithology	Local aquifers: Maryland	Local aquifers: Delaware	Aquifer and confining unit used in this report: Model layer number	Aquifer and confining unit used in this report: Model name
Quaternary	Holocene to Pliocene	Columbia Group (undivided)	Columbia Formation (undivided)	Sand, mostly coarse; moderately sorted with gravel and occasional cobbles and thin silt layers.	Columbia	Columbia	10	Surficial aquifer
Tertiary	Miocene	Chesapeake Group[3]: "Yorktown and Cohansey[2] and Eastover Formations"	Chesapeake Group (undivided)	Sand, interbedded gray to whitish gray; fine to coarse grained, and dark gray to blue-gray clay and silt.			9	Upper Chesapeake confining unit
					Pocomoke	Pocomoke		
					Ocean City			Upper Chesapeake aquifer
					Manokin	Manokin		
		St. Marys(?) Formation		Clay, gray, clayey silt, and very fine sand. Often fossiliferous. Coarsens upwards.			8	St. Marys confining unit
		Choptank Formation		Silty sand to sand; medium to coarse grained, with some gravel and locally abundant shells. Interbedded with gray to bluish-gray, sandy silt and clay. Becomes predominantly silty west of the Chesapeake Bay.	Frederica	Frederica		Lower Chesapeake aquifer
					Federalsburg	Federalsburg		
		Calvert Formation			Cheswold	Cheswold		
	Eocene	Piney Point and Nanjemoy Formations	Piney Point and Nanjemoy Formations	Sand, grayish-green to grayish-white; medium to coarse grained; glauconitic calcite cemented layers. Shell debris. Does not crop out in Maryland-Delaware. Coarsens upwards from basal silt.	Piney Point	Piney Point	7	Lower Chesapeake confining unit; Piney Point–Nanjemoy Aquifer
	—?—	Marlboro Clay	Rancocas Group	Clay, silty, reddish brown to pink or gray; micaceous.			6	Nanjemoy-Marlboro confining unit
	Paleocene	Aquia Formation		Sand, very fine to coarse-grained; poorly to well sorted, yellow, purple, or green colored, with glauconite, lignite, and shell material. Clay layers are greenish gray to black with glauconite.	Aquia	Rancocas		Aquia-Rancocas aquifer
		Brightseat Formation					5	Lower Brightseat confining unit

freshwater interface (designated as the 10,000-mg/L (milligrams per liter) isochlor; Meisler, 1981, fig. 4). Meisler and others (1984, p. 8) tested the use of a no-flow boundary at the interface and successfully applied this method to simulate ground-water flow in the Coastal Plain of southern New Jersey. Overlying the aquifer is a thick sequence of silt and clay of the Arundel Formation. Underlying the aquifer is the consolidated rock basement complex. The total area of the aquifer is 8,217 mi^2, or about 63 percent of the modeled area. The top of the Patuxent aquifer defines a surface that dips about 0.5° southeast from its outcrop near the Fall Line. Along the eastern boundary, as defined by the 10,000-mg/L isochlor, the top of the aquifer ranges from 1,800 ft below sea level at the New Jersey line to 2,900 ft below sea level along the southern part of the Eastern Shore. The aquifer ranges in thickness from a featheredge at the Fall Line to 360 ft on the Eastern Shore (Mack, 1983, fig. 4).

TABLE 1.—*Relation of stratigraphic units, hydrogeologic units, and equivalent layers in the computer model of the Coastal Plain of Maryland and Delaware (Vroblesky and Fleck, 1991, plate 1)*—Continued

System	Series	Stratigraphic units [1]: Maryland	Stratigraphic units [1]: Delaware	Lithology	Local aquifers: Maryland	Local aquifers: Delaware	Aquifer and confining unit used in this report: Model layer number	Aquifer and confining unit used in this report: Model name
Cretaceous	Upper Cretaceous	Severn Formation	Monmouth Formation	Sand, fine- to coarse-grained; silty or clayey, reddish brown; glauconitic. Localized occurrences of poorly sorted coarse-grained sand.	Severn	Severn	5	Severn aquifer
							4	Severn confining unit
		Matawan Formation	Matawan Group	Sand, fine-grained; silty or clayey; dark gray; micaceous; glauconitic.	Matawan	Matawan	4	Matawan aquifer
							3	Matawan confining unit
		Magothy Formation	Magothy Formation	Sand, loose, fine- to coarse-grained; white, associated with lignite and dark laminated silt-clay sand and gravel; coarser near base. Top grades texturally into clay.	Magothy	Magothy	3	Magothy aquifer
		Potomac Group: Raritan and Patapsco Formation	Potomac Formation	Sand, fine- to medium-grained; interbedded with variegated (red to gray) silt or clay. Abrupt lateral and vertical changes in lithology. Predominantly sandy in the north, becoming increasingly silty toward the south.	Lexington Park[4]		2	Patapsco confining unit
	Upper Cretaceous / Lower Cretaceous				Patapsco	Upper and middle hydrologic zones[5]	2	Patapsco aquifer
	Lower Cretaceous	Potomac Group: Arundel Formation	Potomac Formation	Clay, thick, variegated; dense with increasing amounts of interbedded sand lenses in a downdip direction from the outcrop zone.			1	Potomac confining unit
		Potomac Group: Patuxent Formation		Sand or pebbly sand and gravel; medium to coarse grained, with abrupt lateral and vertical changes to variegated silt and clay.	Patuxent	Lower hydrologic zone	1	Patuxent aquifer
Jurassic (?) to Precambrian		Basement rock	Basement rock	Schists, granites, gneisses, and gabbros.	—	—		

[1] Modified from Jordan and Smith (1983).
[2] Rasmussen and Slaughter (1955).
[3] Since the preparation of this report, Upper Oligocene beds have been assigned to the base of the Chesapeake Group.
[4] Lexington Park aquifer exists only in the southernmost part of the model and is equivalent to layer 3 of the Virginia RASA model, and is overlain by the Lexington Park confining unit (Meng and Harsh, 1984). The small part that occurs in Maryland is therefore included in layer 3 of this study (pl. 1*A*).
[5] Sundstrom and others (1967).

Reported transmissivity values for the Patuxent aquifer range between 1,300 and 11,000 ft^2/d (feet squared per day) (Hansen, 1972, p. 20). Highest transmissivities are located in an area south of Baltimore. Transmissivity is best defined along the western edge of the aquifer. The average storage coefficient for the aquifer near Baltimore is reported to be 2.6×10^{-4} (Bennett and Meyer, 1952, p. 50). Mack (1962, table 4) reports storage coefficients that range from 1×10^{-5} to 1×10^{-3}. Pumpage from this aquifer in 1980 was about 56 ft^3/s (cubic feet per second). Total pumpage from the Coastal Plain aquifers of Maryland and Delaware in 1980 was about 346 ft^3/s (Wheeler and Wilde, 1989, p. 6; Hodges, 1985, p. 167; Cushing and others, 1973, p. 51). Thus, pumpage from the Patuxent aquifer represents about 16 percent of all pumpage from the study area.

Head distribution for 1980, as shown on plate 1*D*, clearly demonstrates the effects of large withdrawals both south of Baltimore and in northern Delaware (Martin and Denver, 1982, table 4). In 1980, pumpage from the area in and around Baltimore totaled about 28 ft^3/s (Wheeler and Wilde, 1989) (pl. 1*E*), resulting in draw-

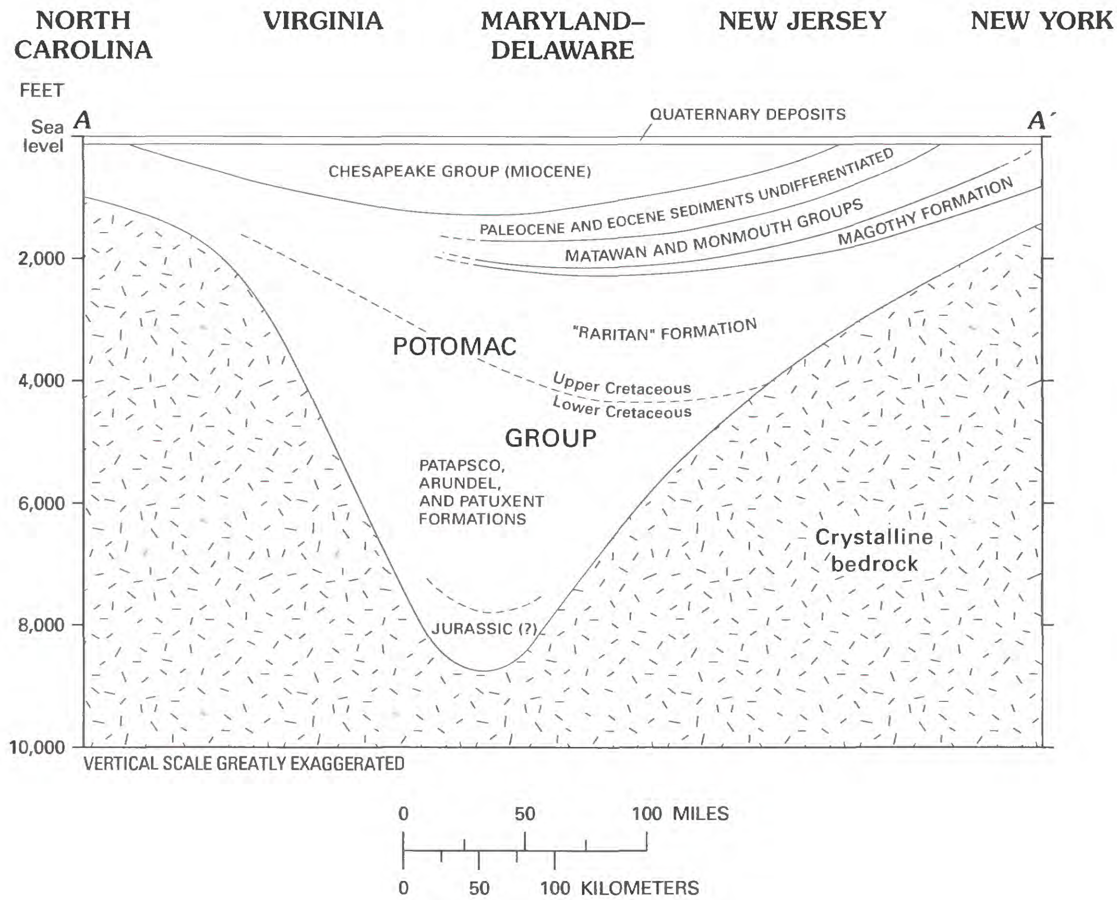


FIGURE 2.—Diagrammatic section of the Salisbury Embayment (modified from Glaser, 1968, fig. 2).

downs of as much as 50 ft. In northern Delaware, pumpage in 1980 amounted to about 19 ft^3/s, resulting in a large cone of depression and drawdowns of about 130 ft (pl. 1*D*). The observation wells shown on plates 1*G*, 1*H*, 1*I*, 1*J*, and 1*K* depict head declines during the past 30 years. The heads in observation well AA Cc 80 (pl. 1*J*) have declined almost 70 ft.

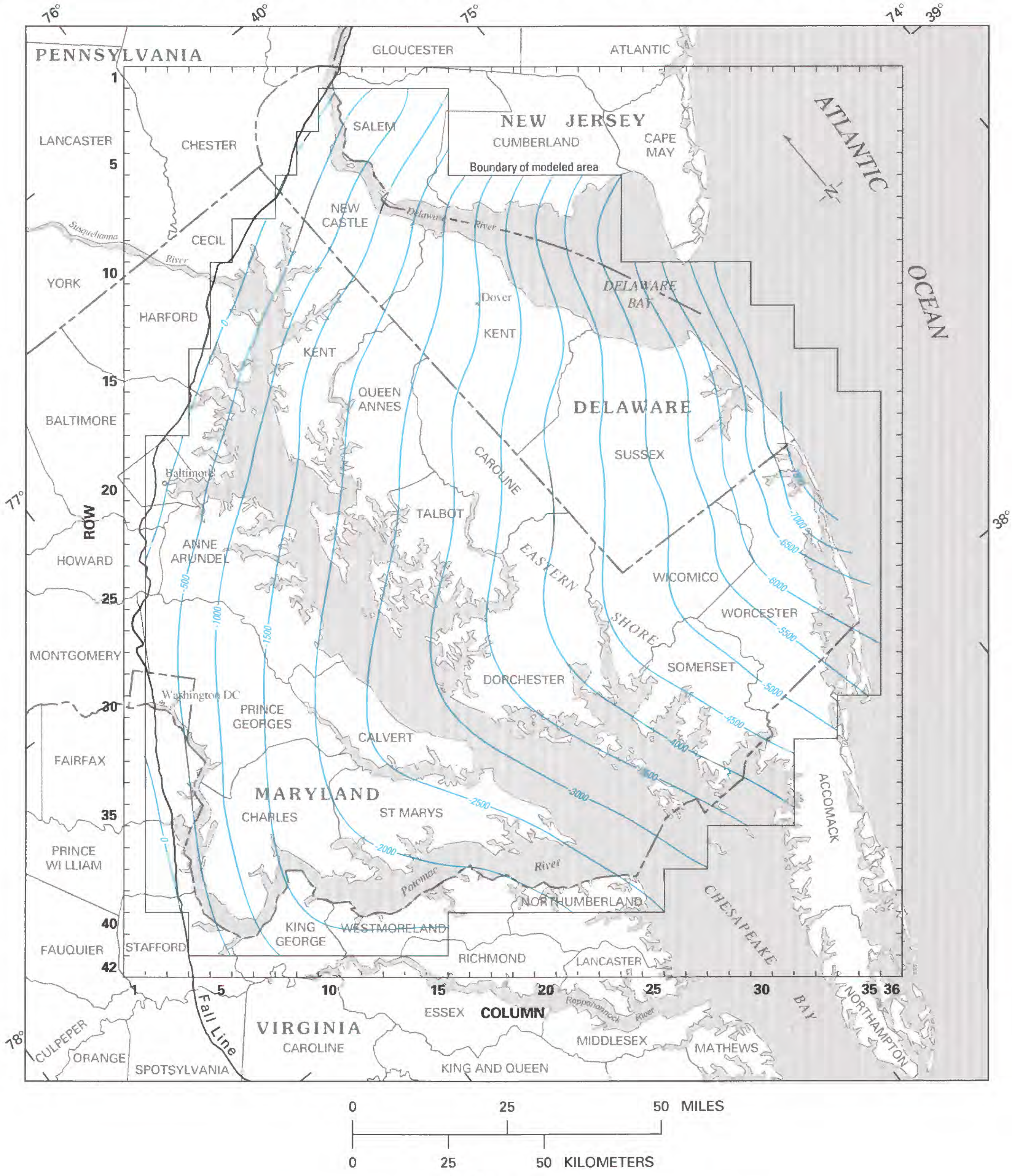
POTOMAC CONFINING UNIT

The Potomac confining unit overlies the Patuxent aquifer in model layer 1. In the updip area, it represents the Arundel Formation, and downdip, it represents the principal clay section between the Patuxent and Patapsco aquifers. The thickness of the unit ranges from a featheredge where it pinches out along the outcrop of the Patuxent aquifer, to as much as 400 to 700 ft along the eastern boundary of the Patuxent aquifer (Vroblesky and Fleck, 1991, fig. 9). No reported values for vertical hydraulic conductivity were available; however, Mack and Achmad (1986, p. 37) obtained a value of 5.9×10^{-7} ft/d (feet per day) from a steady-state calibration of a digital model of the Potomac aquifers in southern Maryland.

LAYER 2

PATAPSCO AQUIFER

The Patapsco aquifer in Maryland is the sandy portion of the Patapsco Formation of the Potomac Group of Early and Late Cretaceous age; in Delaware, it is equivalent to the middle and upper hydrologic zones of the Potomac Formation as described by Martin and Denver (1982, fig. 4). The Patapsco is a multi-aquifer system; however, in this study it is combined into one unit and is referred to as the “Patapsco aquifer.” The extent of the aquifer in the modeled area is shown on plate 2*A*. The aquifer is bounded on the west by the outcrop of the Patapsco Formation, the northern and



EXPLANATION

1000 **Subsurface contour**—Shows altitude of top of basement rock. Contour interval 500 feet. Datum is sea level

FIGURE 3.—Structural surface of the top of the basement rocks (modified from Brown and others, 1972).

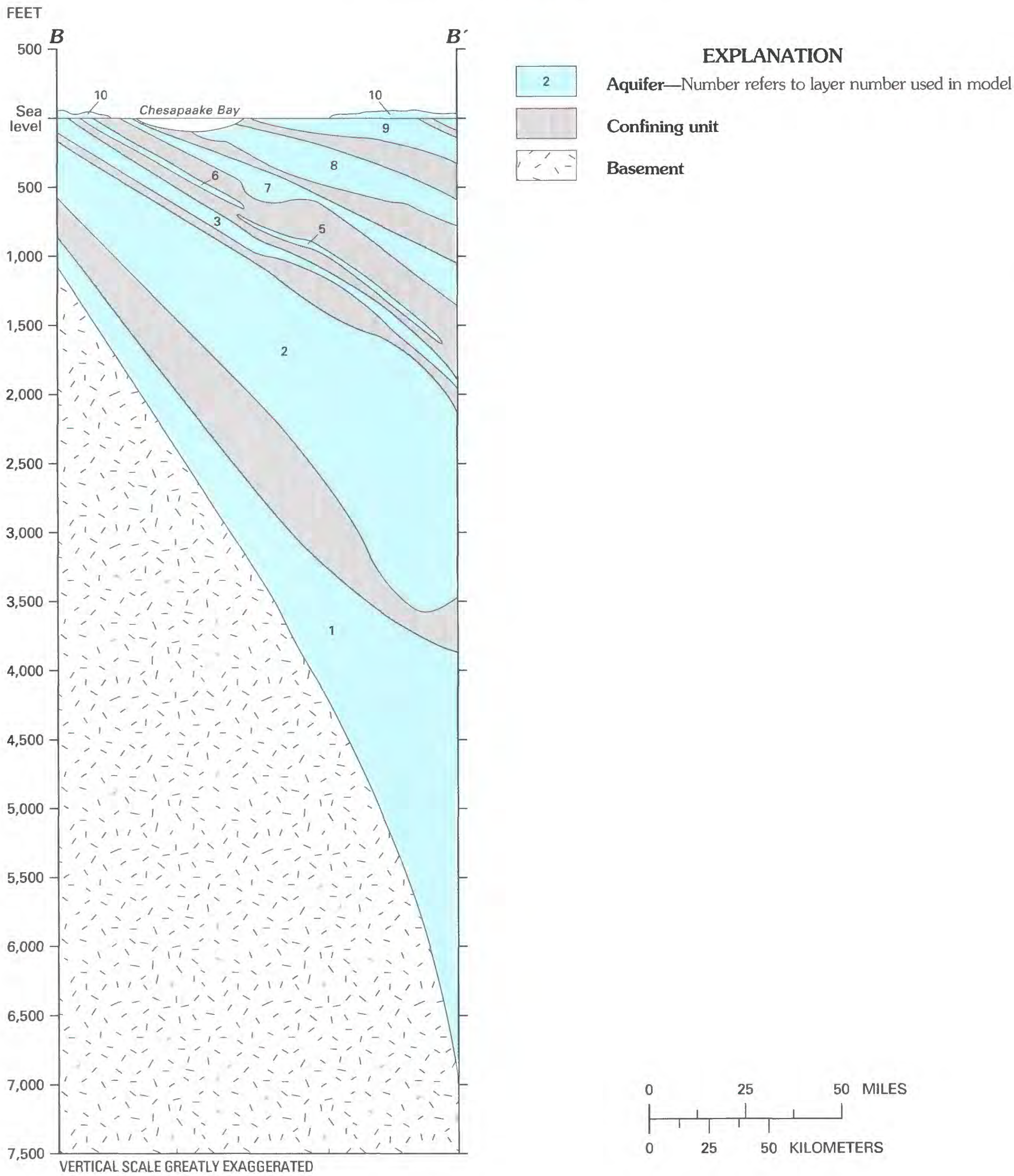


FIGURE 4.—Diagrammatic section of major hydrogeologic units (modified from Meisler, 1986, fig. 104).

southern boundaries are flux boundaries at the New Jersey and Virginia State lines, and the eastern no-flow boundary is defined by the 10,000-mg/L isochlor (Meisler, 1981, fig. 4). The modeled area of the aquifer is about 10,775 mi^2, which is about 82 percent of the total area of the model. Professional Paper 1404–E (Vroblesky and Fleck, 1991, p. 13) gives a complete description of the hydrogeologic framework of the aquifer as developed for this study.

The Patapsco aquifer is most transmissive in an area south of Baltimore, where the transmissivity of individual sand units is as high as 9,000 ft^2/d (Papadopulus and others, 1974, fig. 3). Northward and southward, the transmissivity of the aquifer decreases. Eastward, there is a paucity of data. However, data from a few deep test wells on the Eastern Shore indicate a fining of the sand, which together with the thickening wedge probably indicates a slight reduction in transmissivity (Hansen, 1972, p. 35). Reported storage coefficients range from 3×10^{-5} (Mack, 1962, table 5) to 5×10^{-3} (Hansen, 1972, p. 34).

The thickness of the Patapsco aquifer varies from a featheredge along the western boundary to as much as 1,020 ft in a deep test well on the Eastern Shore of Maryland (Trapp and others, 1984, table 12). Typically, in southern Maryland the thickness of the aquifer ranges from 500 to 900 ft (Hansen, 1972, pl. 2). The thickness of the sand lenses within the aquifer varies considerably. In the area of high transmissivity, the sand probably accounts for as much as 75 percent of the total thickness; whereas elsewhere, the sand thickness may be no more than 50 percent.

The Patapsco aquifer is heavily pumped. In 1980, about 52 ft^3/s was pumped from the aquifer throughout the modeled area (Wheeler and Wilde, 1989; Hodges, 1985, p. 167; Cushing and others, 1973, p. 51). The greatest proportion of this withdrawal is from an area just south of Baltimore, where about 31 ft^3/s was withdrawn in 1980. In northern Delaware, pumpage from the aquifer in 1980 amounted to about 12 ft^3/s (Martin and Denver, 1982, table 3). Pumpage from the aquifer in the Baltimore industrial area reached a peak of about 12 ft^3/s during the 1940's. However, due to intrusion of brackish water, pumpage in this area has decreased to less than 1 ft^3/s (Wheeler and Wilde, 1989). Total pumpage from the aquifer during 1930–80 increased about eightfold.

Increasingly heavy pumpage in the Patapsco aquifer has caused lowering of heads and development of cones of depression, especially in the area south of Baltimore and in northern Delaware. Observation well AA Cc 82 (pl. *2L*) is typical of this trend and indicates that during the 20-year period from 1961 to 1980, there was about 20 ft of drawdown.

PATAPSCO CONFINING UNIT

The Patapsco confining unit represents the upper clay facies of the Patapsco Formation. Where the Magothy Formation is present, its base defines the top of the unit. The extent of the unit is defined by the confined part of the Patapsco aquifer. The thickness of the unit varies from a featheredge along the outcrop of the Patapsco aquifer to a maximum of 400 ft (Vroblesky and Fleck, 1991, fig. 11) in eastern Delaware at the limit of the Patapsco aquifer as defined by the 10,000-mg/L isochlor. In southern Maryland, the thickness of the unit averages about 50 ft, never exceeding 100 ft (Hansen, 1968, pls. 2–17). Of particular importance in the control of the flow pattern within the Patapsco aquifer is the extensive erosion of the confining unit by the ancestral Chesapeake Bay. This erosion has resulted in a considerable thinning of the unit underlying the bay. Reported values for vertical hydraulic conductivity range from 5.9×10^{-7} to 1.08×10^{-5} ft/d in southern Maryland (Mack and Mandle, 1977, table 2) and an estimated value of about 2×10^{-6} ft/d for Delaware (Phillips, 1987, table 2).

LAYER 3

MAGOTHY AQUIFER

The aquifer in the central and northern part of layer 3 (pl. *3A*) is equivalent to the sand section of the Upper Cretaceous Magothy Formation. In the extreme southern part of Maryland, the modeled layer represents the upper Potomac aquifer (Trapp, 1992). The sediment of the upper Potomac aquifer is also Cretaceous in age but is not in direct hydrologic contact with sediment of the Magothy Formation. The two aquifers are herein referred to as the “Magothy aquifer.” The modeled aquifer area is about 68 percent of the modeled area, or approximately 8,920 mi^2. The areal extent of the Magothy aquifer is bounded on the west by the outcrop area, on the north by the New Jersey State line, and on the east by the assumed no-flow boundary at the 10,000-mg/L isochlor (Meisler, 1981, fig. 4). The part of the aquifer corresponding to the Magothy Formation pinches out in southern Maryland. The part of the aquifer representing the upper Potomac aquifer extends southward into Virginia.

The Magothy aquifer is most transmissive in an area south of Baltimore, where transmissivity values of about 8,000 to 12,000 ft^2/d are obtained (Hansen, 1972, fig. 15). Throughout the rest of the modeled area, transmissivities generally range from 1,000 to 3,000 ft^2/d. Reported storage coefficients range from 7×10^{-5} to 3×10^{-4} (Mack, 1962, table 6).

The thickness of the Magothy aquifer ranges from a featheredge, where it pinches out to the south and along the outcrop, to about 200 ft in an area south of Baltimore (Hansen, 1972, pl. 3). Typically, the thickness of the aquifer ranges from 50 to 100 ft.

In 1980, total pumpage from the Magothy aquifer was about 29 ft^3/s (Wheeler and Wilde, 1989; Hodges, 1985, p. 167; Cushing and others, 1973, p. 51), about 80 percent of which was from the areas of high transmissivity shown on plate 3*A*. Pumpage increased from about 0.5 ft^3/s in 1930 to about 29 ft^3/s in 1980 (Wheeler and Wilde, 1989). This large increase is reflected by the hydrograph of plate 3*Q*, where heads have decreased in southern Maryland as much as 70 ft since 1962, to an altitude of about 35 ft below sea level.

Over the past 50 years, increased pumpage has resulted in an altered pattern of head distribution, as indicated on plate 3*D*. Especially noteworthy are the two large cones of depression that have developed southeast of Washington, D.C., causing a reversal of head gradients.

MATAWAN CONFINING UNIT

The part of the Magothy aquifer corresponding to upper Potomac sands is overlain by the Brightseat confining unit (Vroblesky and Fleck, 1991, pl. 1). The part of the Magothy aquifer representing the Magothy Formation is overlain by the Matawan confining unit and is composed of the clay-silt facies of the Matawan Formation that reaches a maximum thickness of about 100 ft across north-central Delaware and southwestward into Maryland. Along the Atlantic Coast of Maryland, the unit attains a thickness of about 80 ft (Vroblesky and Fleck, 1991, fig. 13). Elsewhere, it thins abruptly, pinching out at its areal limits. Analyses of samples of the Matawan confining unit from two sites in southern Maryland indicate a range of vertical hydraulic conductivity from 5.7×10^{-5} to 3.1×10^{-4} ft/d (Mack and Mandle, 1977, table 2).

LAYER 4

MATAWAN AQUIFER

The layer 4 (pl. 4*A*) aquifer represents the sandy part of the Matawan Formation in the model. In the New Jersey section of the northern Atlantic Coastal Plain RASA study, layer 4 represents the Englishtown Formation, which is a productive aquifer. However, in Maryland and Delaware, layer 4 generally is not an aquifer. Where layer 4 does exist as an aquifer, it is silty and provides only limited amounts of water. In the modeled area, the Matawan aquifer encompasses an area of only about 1,071 mi^2, or 8 percent of the modeled area.

For this study, the boundaries of the Matawan aquifer are defined to include the few known occurrences of pumpage from this aquifer. The aquifer extends from the north, where it matches with the Englishtown aquifer as defined in the New Jersey area, and southward through parts of northern and central Delaware into a small area of Maryland.

Transmissivity of the Matawan aquifer, as determined from specific capacity data (Mack and others, 1971, table 16), reaches a maximum of about 2,500 ft^2/d on the central Eastern Shore of Maryland. Elsewhere, transmissivity is low, never exceeding 1,000 ft^2/d. No values for storage coefficients for the Matawan aquifer are reported.

Pumpage from the Matawan aquifer is negligible. There is no major pumpage at all and only a small amount of domestic pumpage in Maryland and central Delaware.

SEVERN CONFINING UNIT

The Severn confining unit is equivalent to the clayey facies of the lower portion of the Severn Formation. The unit overlies the sandy facies of the Matawan Formation, and together these units are modeled as layer 4.

The Severn confining unit is generally a clayey silt, and some mica and glauconite are present. It averages about 40 ft in thickness (Hansen, 1968, pls. 2–17), reaching a maximum of about 90 ft in the western part of Delaware. No data for vertical hydraulic conductivity were available for the Severn confining unit. Because of the silty nature of the unit, it is estimated that the vertical hydraulic conductivity is about 3×10^{-4} ft/d.

LAYER 5

SEVERN AQUIFER

The aquifer of layer 5 (pl. 4*G*) represents that part of the Severn and Monmouth Formations that is sufficiently sandy to function as an aquifer. The Severn aquifer encompasses an area of 5,560 mi^2, or about 42 percent of the modeled area. This area extends from New Jersey southwestward through northern Delaware, across the upper Eastern Shore of Maryland, across the Chesapeake Bay into a small portion of southern Maryland, and southward across southern Delaware and the lower Eastern Shore of Maryland. Elsewhere, the Severn Formation either has been eroded away or acts as a confining unit.

The top of the Severn aquifer dips southeasterly from the outcrop along its northwest boundary toward the Delaware coastline. The lowest altitude of the top of the aquifer is about 2,100 ft below sea level in the vicinity of the Atlantic Coast and the Maryland-Delaware line.

The thickness of the Severn aquifer varies from as little as 5 ft to a maximum of about 150 ft in central Delaware (Hansen, 1968, pl. 6). In northern and central Delaware and into a small part of Maryland, the thickness of the aquifer averages about 90 ft. Reported transmissivity values range from about 200 to 800 ft^2/d (Hansen, 1972, p. 110). Storage coefficients are reported to range from 3×10^{-7} to 3×10^{-4} (Hansen, 1972, p. 110).

Pumpage from the Severn aquifer amounts to about 0.4 ft^3/s (Wheeler and Wilde, 1989). More than 90 percent of this total is withdrawn from two small areas of Maryland (pl. 4*K*); the remaining pumpage occurs in Delaware.

LOWER BRIGHTSEAT CONFINING UNIT

The Brightseat Formation in Maryland and the clayey facies of the Rancocas Group in Delaware are represented in the model by the lower Brightseat confining unit. This unit overlies the sandy facies of the Severn Formation and, together with the Severn aquifer, represents layer 5 of the model. The composition of this part of the Rancocas, much like the clayey facies of the Severn Formation, is principally a dark micaceous and glauconitic, clayey silt. The areal extent of the unit corresponds to the Severn aquifer, except where the latter subcrops or crops out. The thickness of this confining unit varies from a featheredge, where it pinches out along the western and southern sides of the model area coincident with the limits of the Severn aquifer, to a maximum thickness of about 200 ft on the eastern side of central Delaware (Vroblesky and Fleck, 1991). The average thickness is about 100 ft. No data for vertical hydraulic conductivity for the lower Brightseat confining unit were available. Because of the clayey nature of this unit, it is estimated that the vertical hydraulic conductivity is about 6×10^{-5} ft/d.

LAYER 6

AQUIA-RANCOCAS AQUIFER

The aquifer of layer 6 is the Aquia Formation in Maryland and is the Rancocas Formation in Delaware. The total area of the Aquia-Rancocas aquifer is about 5,120 mi^2, or approximately 39 percent of the modeled area. The extent of the aquifer is shown on plate 5*A*. The aquifer is defined on the west by the outcrop (or subcrop), which within the model, encompasses an area of 618 mi^2. The eastern boundary of the aquifer is defined by a facies change in the Aquia Formation from sand to silt (Chapelle and Drummond, 1983, p. 40). The northern and southern boundaries are the boundaries of the model area. The aquifer dips southeastward from its outcrop. The maximum depth to the top of the aquifer occurs on the Eastern Shore of Maryland at an altitude of about 700 ft below sea level.

The Aquia-Rancocas aquifer varies in thickness from a featheredge on the west to 250 ft southeast of Baltimore and across the Chesapeake Bay into western Delaware (Vroblesky and Fleck, 1991, pl. 1). There, the thickness of the aquifer averages about 200 ft.

The transmissivity of the Aquia-Rancocas aquifer is controlled partly by the thickness and partly by the gradual eastward facies change from sand to silt. Transmissivity distribution of the aquifer, as determined from available data (Hansen, 1974, fig. 9; Chapelle and Drummond, 1983, fig. 5), increases eastward from a maximum of 2,000 ft^2/d in southern Maryland to as much as 5,100 ft^2/d on the Eastern Shore. Reported values of storage coefficient range from 1×10^{-4} to 4×10^{-4} (Hansen, 1972, p. 66).

Pumpage from the aquifer by major users increased from about 0.6 ft^3/s in 1940 to about 9 ft^3/s in 1980 (Wheeler and Wilde, 1989). The greatest amount of pumpage by major users occurs near the southern tip of Maryland (pl. 5*E*), where about 1.1 ft^3/s were withdrawn in 1980. The Aquia aquifer is heavily pumped by small users. Total pumpage from the aquifer for 1980 was about 28 ft^3/s (Wheeler and Wilde, 1989; Hodges, 1985, p. 167; Cushing and others, 1973, p. 51). In 1980, pumpage averaged about 11 ft^3/s on the Eastern Shore and about 17 ft^3/s in southern Maryland. Because of the high transmissivity of the aquifer on the Eastern Shore, drawdown due to pumpage has been negligible. Near the southern tip of Maryland, transmissivity is only about 900 ft^2/d; thus, heavy pumpage has resulted in drawdowns of up to 75 ft (pl. 5*D*).

NANJEMOY-MARLBORO CONFINING UNIT

The Nanjemoy-Marlboro confining unit overlying the Aquia-Rancocas aquifer represents the Marlboro Clay and the lower section of the Nanjemoy Formation. The unit also includes much of the Aquia Formation in southern Maryland and the lower Eastern Shore, where the Aquia Formation undergoes a facies change. The average thickness of the unit is about 100 ft on the western side of the Chesapeake Bay and about 250 ft on the Eastern Shore, reaching a maximum thickness of about 400 ft in southwestern Delaware (Vroblesky and Fleck, 1991, fig. 21).

Vertical hydraulic conductivities from several locations in southern Maryland range from 6×10^{-5} to 6×10^{-4} ft/d (Chapelle and Drummond, 1983, table 2). The lower end of this range probably is a better representation of the conductivity of the tight Marlboro Clay; therefore, a value of 6×10^{-5} ft/d was used as the initial value in the early model simulations.

LAYER 7

PINEY POINT–NANJEMOY AQUIFER

Above layer 6 is the Piney Point Formation, which extends throughout much of the study area. The Piney Point Formation generally is a productive aquifer and, in this study, is represented by the Piney Point–Nanjemoy aquifer. Where the Nanjemoy Formation, which underlies the Piney Point Formation, is sufficiently sandy to act as an aquifer, it is included as part of the aquifer. Within the modeled area, the aquifer encompasses an area of 6,670 mi^2, or about 51 percent of the modeled area. The eastern and western boundaries, as indicated on plate 6*A*, are defined by the Piney Point Formation subsurface pinch-out. To the north and south, the limits of the aquifer are coincident with limits of the model.

The Piney Point–Nanjemoy aquifer dips southeastward from its western limit toward southern Delaware. Structure contours on the top of the aquifer are about 40 ft above sea level in southern Maryland (Vroblesky and Fleck, 1991, fig. 22). At the eastern limit in southern Delaware, the top of the aquifer is 1,000 ft below sea level.

Two distinct transmissivity highs occur within the study area (Williams, 1979, pl. 5; Leahy, 1979, fig. 16). One high is centered in central Delaware, where the Piney Point–Nanjemoy aquifer is sandy and the thickness ranges from about 200 to 250 ft. Transmissivity in central Delaware is as much as 6,400 ft^2/d. Downdip in southern Delaware, the formation gradually becomes siltier with a concomitant reduction in transmissivity. Southward, in coastal Maryland, the Piney Point Formation ceases to exist as an aquifer.

A second area of high transmissivity occurs in the central part of Maryland's Eastern Shore. In this area, the Piney Point–Nanjemoy aquifer reaches a maximum thickness of about 190 ft (Trapp and others, 1984, table 2), and transmissivity values are as high as 5,000 ft^2/d. Elsewhere, the aquifer is thinner and siltier, resulting in lower transmissivity.

Reported storage coefficients for the Piney Point aquifer range from 9×10^{-5} to 4×10^{-4} (Leahy, 1979, p. 17; Williams, 1979, p. 13; Hansen, 1972, p. 80).

Two major cones of depression have developed in the Piney Point–Nanjemoy aquifer, as indicated on plate 6*D*. One cone is in the vicinity of Dover, Del., and the second cone is on the Eastern Shore of Maryland. Total pumpage from the aquifer for 1980 (pl. 6*E*) averaged slightly more than 14 ft^3/s (Wheeler and Wilde, 1989; Hodges, 1985, p. 167; Cushing and others, 1973, p. 51). In central Delaware, pumpage of 6.4 ft^3/s in 1980 caused drawdowns of as much as 150 ft (pl. 6*J*). However, on the Eastern Shore in the area of observation well DO Ce 21 (pl. 6*J*), pumpage decreased from 5.9 ft^3/s in 1970 to only 2.0 ft^3/s in 1980. In 1970, heads were 100 ft below sea level, but by 1980, a recovery of about 30 ft had occurred.

LOWER CHESAPEAKE CONFINING UNIT

The lower Chesapeake confining unit is above the Piney Point–Nanjemoy aquifer, and together they form layer 7 in the model. This unit represents the clay facies of the Piney Point and Calvert Formations. Typically, this unit is silty and contains some clay and silty sand of low permeability. The lower Chesapeake confining unit reaches a maximum thickness of about 240 ft and has an average thickness of 100 ft (Leahy, 1979, p. 18).

Vertical hydraulic conductivities of the lower Chesapeake confining unit in the vicinity of Dover, Del., range from 4×10^{-5} to 9×10^{-5} ft/d (Leahy, 1979, p. 28) and in Maryland from 7.3×10^{-5} to 1×10^{-2} (Chapelle and Drummond, 1983, table 2).

LAYER 8

LOWER CHESAPEAKE AQUIFER

Layer 8 in the study area is equivalent to a major part of the lower section of the Chesapeake Group and encompasses an area of about 6,940 mi^2, or approximately 55 percent of the modeled area. The lower section of the Chesapeake Group typically includes three water-bearing units: the Cheswold, Federalsburg, and Frederica aquifers. In this study, these aquifers together represent the lower Chesapeake aquifer (pl. 7*A*).

The lower Chesapeake aquifer extends from the outcrop of the Cheswold-Calvert unit across central Delaware into Maryland and south across the rest of the Eastern Shore. Structure contours of the top of the aquifer as mapped in Professional Paper 1404–E (Vroblesky and Fleck, 1991, fig. 24) indicate that the aquifer dips to about 700 ft below sea level along the Maryland Atlantic Coast. Although some lower Chesapeake sediments occur in southern Maryland, they do not function as an aquifer in that area.

Typically, the lower Chesapeake aquifer is about 300 ft thick (Cushing and others, 1973, p. 43–45). However, since the aquifer consists of one or more confining units, the average hydraulic conductivity generally is not more than 10 to 15 ft/d. Reported transmissivity values of the individual aquifers range from 350 to 7,400 ft^2/d for the Cheswold aquifer (Leahy, 1982a, p. 16), 450 to 1,400 ft^2/d for the Federalsburg aquifer (Cushing and others, 1973,

p. 44), and up to 1,400 ft^2/d for the Frederica aquifer (Cushing and others, 1973, p. 45). In the vicinity of Dover, Del., the maximum transmissivity of the Cheswold aquifer is about 7,400 ft^2/d (Leahy, 1982a, p. 16). Storage coefficients range from 1×10^{-4} in the Federalsburg aquifer (Cushing and others, 1973, p. 44) to 6×10^{-3} in the Cheswold aquifer (Leahy, 1982a, p. 19).

Total pumpage from the lower Chesapeake aquifer in the study area was about 26 ft^3/s in 1980 (Wheeler and Wilde, 1989; Hodges, 1985, p. 167; Cushing and others, 1973, p. 51). Plate *7E* indicates that the largest pumpage was concentrated in the vicinity of Dover. Drawdown in this area by 1977 was about 100 ft (pl. *7D*).

ST. MARYS CONFINING UNIT

The St. Marys confining unit represents the St. Marys Formation in Maryland and the clay and silt sequence in the middle of the Chesapeake Group in Delaware. Within the study area, the unit overlies the aquifers that are represented by the lower Chesapeake aquifer. Thus, layer 8 of the model represents the lower and middle parts of the Chesapeake Group. The unit consists typically of gray clay, silt, and very fine sand, and tends to coarsen upwards. The thickness of the unit as used in the model varies from a featheredge, where it pinches out along the subcrop of the lower Chesapeake aquifer, to about 240 ft on the Eastern Shore of Maryland (Vroblesky and Fleck, 1991, fig. 25), and averages about 100 ft.

No vertical hydraulic conductivity data were available. Because of the sandy nature of this unit, a relatively high value of 4×10^{-4} ft/d was initially used in the flow model.

LAYER 9

UPPER CHESAPEAKE AQUIFER

To be consistent with the overall northern Atlantic Coastal Plain RASA model, layer 9 represents the upper Chesapeake Group, consisting of three water-bearing sand units. These units are the Manokin, Ocean City, and Pocomoke aquifers and herein are referred to as the "upper Chesapeake aquifer." The area of the aquifer is about 4,690 mi^2, or 36 percent of the modeled area (pl. *8A*).

The upper Chesapeake aquifer within the study area is a shallow aquifer. The top of the unit is about 150 ft below sea level along the Atlantic Coast in Maryland. Thickness of the aquifer ranges from a featheredge along the outcrop to as much as 350 ft at its southeastern extremity (Cushing and others, 1973, pls. 9–10). Reported transmissivity values range from about 1,000 to 20,000 ft^2/d for the Ocean City–Manokin aquifer system (Weigle, 1974, p. 18) and 1,000 to 8,000 ft^2/d for the Pocomoke aquifer (Cushing and others, 1973, p. 46). Storage coefficients range from 1×10^{-4} to 3×10^{-3} for the upper Chesapeake aquifer (Cushing and others, 1973, p. 45–46).

Total pumpage from the aquifer amounted to 19 ft^3/s in 1980 (Wheeler and Wilde, 1989; Hodges, 1985, p. 167; Cushing and others, 1973, p. 51), of which about 12 ft^3/s was from the area along the Atlantic Coast of Maryland (pl. *8E*). From 1940 to 1980, pumpage increased about eightfold, which resulted in several local cones of depression (Weigle and Achmad, 1982, p. 1).

UPPER CHESAPEAKE CONFINING UNIT

The upper Chesapeake confining unit overlying the upper Chesapeake aquifer is a discontinuous unit of lenticular silt, clay, and fine sand. This unit separates the upper Chesapeake aquifer from the Pliocene-Pleistocene sediments represented in the model by the surficial aquifer. The extent of the unit in the study area is limited to the extreme southeastern portion of the model. The average thickness of the unit is about 70 ft, reaching a maximum of 180 ft on the lower Eastern Shore of Maryland (Vroblesky and Fleck, 1991, fig. 27).

Because of a lack of data and the sandy and discontinuous nature of the upper Chesapeake confining unit, an initial value of 3.6×10^{-3} ft/d for vertical hydraulic conductivity was entered into the flow model. This value is consistent with a median value of 3×10^{-3} ft/d for 14 samples from Holocene sediments in northern Delaware (Phillips, 1987, table 2).

LAYER 10: SURFICIAL AQUIFER

Layer 10 corresponds to the Pliocene-Pleistocene Columbia water-table aquifer, herein referred to as the "surficial aquifer." The aquifer effectively is two hydrologic units separated by the Chesapeake Bay. The area of the aquifer on the Eastern Shore is 5,250 mi^2, and in southern Maryland is 2,030 mi^2, comprising a total area of 7,280 mi^2, or about 56 percent of the modeled area. The Columbia aquifer in northern Delaware was not included in the surficial aquifer because of its localized and discontinuous extent.

In general, the surficial aquifer is less than 100 ft thick (Johnston, 1973, fig. 3; Bachman and Wilson, 1984, pl. 5). However, on the Eastern Shore there are several well-defined paleochannels (Bachman and Wilson, 1984, p. 11) in which the aquifer thickens to as much as 230 ft. Reported transmissivity values range from 4,000 to 53,000 ft^2/d for the Eastern Shore (Hansen, 1972, p. 117) and from 1,200 to 22,000 ft^2/d for Delaware (Johnston, 1973, p. 12–31). Cushing and others (1973, p. 47) report storage coefficients of 1×10^{-4} to 1.7×10^{-1}. However,

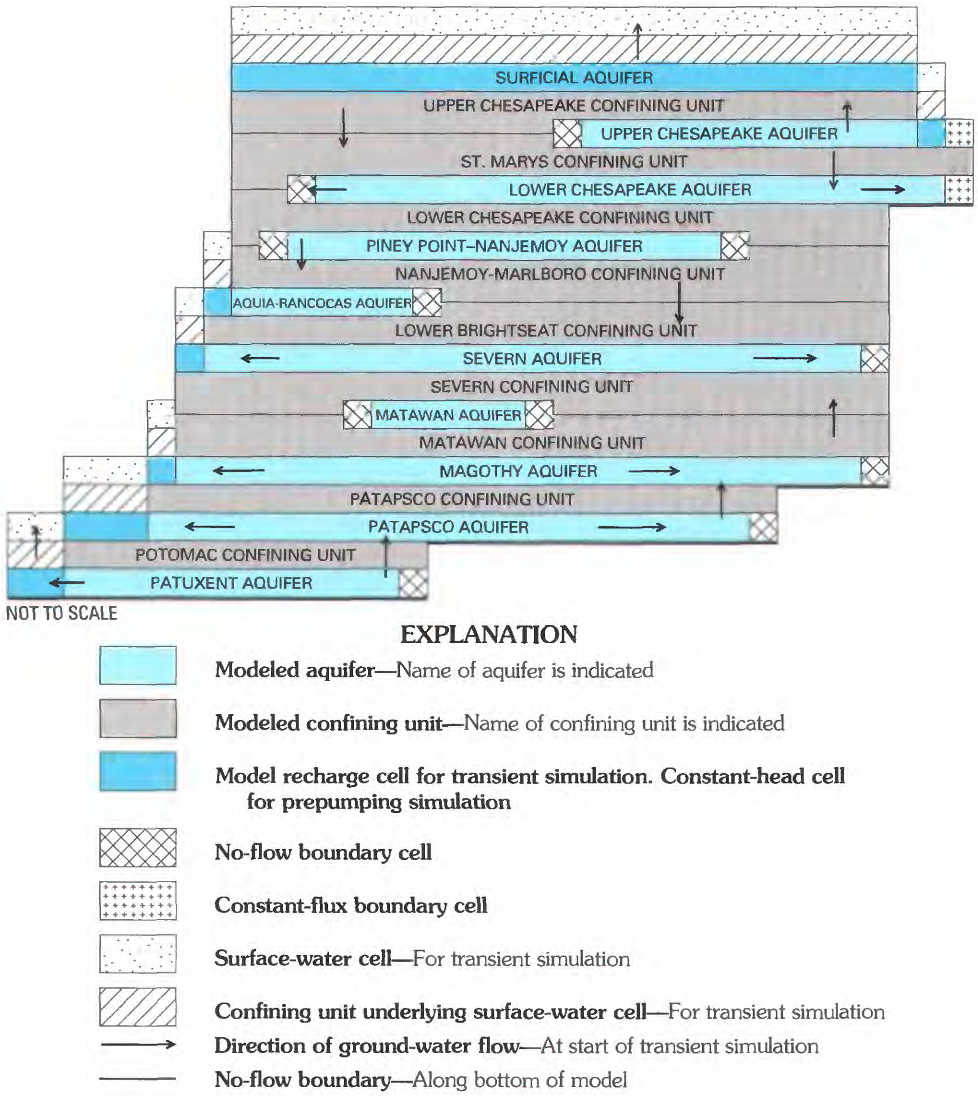


FIGURE 5.—Diagrammatic section showing the conceptual model along model row 20.

average storage coefficients are probably about 1.4×10^{-1} (Johnston, 1973, p. 31).

Total pumpage in the aquifer was about 122 ft^3/s in 1980 (Wheeler and Wilde, 1989; Hodges, 1985, p. 167; Cushing and others, 1973, p. 51). Because of the very high transmissivities and storage coefficients of the surficial aquifer and the abundant recharge from precipitation, drawdown from pumpage has been negligible.

CONCEPTUAL MODEL

The digital model for ground-water flow described in this report is a mathematical representation of the flow system of the Coastal Plain sediments of Maryland, Delaware, and the District of Columbia. The model was constructed by first quantifying the physical characteristics of the flow system and then translating these characteristics into a form that could be manipulated by digital computer. The flow system of the Coastal Plain sediments is complex; therefore, the digital flow model is an idealized and simplified representation of the natural system. This simplified version of the natural system is the conceptual model.

Figure 5 shows diagrammatically the conceptual model that was used to construct the digital model as described in the next section of this report. This model is based on the geologic and hydrologic information discussed in the previous section.

The Coastal Plain sediments within the project area are conceived to consist of 10 aquifers and 9 intervening confining units as described in table 1 and shown in figure 5. The aquifers, in ascending order, are the Patuxent, Patapsco, Magothy, Matawan, Severn, Aquia-Rancocas, Piney Point–Nanjemoy, lower Chesapeake, upper Chesapeake, and surficial. The layers in the model that represent the Patuxent, Patapsco, lower Chesapeake, and upper Chesapeake aquifers are simplifications of actual sediment sequences that include several separate water-bearing units.

The aquifer system is bounded laterally along the west and the bottom by no-flow boundaries. The bottom boundary corresponds to the sloping contact between the bottommost sediments of the Patuxent aquifer and the underlying, nearly impermeable consolidated rocks of mostly Mesozoic age. The Fall Line represents an irregular no-flow boundary at the updip limit of the aquifer system to the west. For the bottom two layers a downdip no-flow boundary to the east represents an assumed saltwater-freshwater interface located where the ground water contains concentrations of chloride of 10,000 mg/L, as delineated by Meisler (1981). The aquifer system to the north and south and for the top eight layers to the east is represented by flux boundaries, specified by the regional model (Leahy and Martin, 1994).

During steady-state simulations, the top boundary had water-table altitudes specified as constant heads. Thus, for steady-state conditions, the water-table aquifer was not simulated; however, the flow recharged from the water-table aquifer to the deeper aquifers was simulated.

Simulation of the flow system for transient conditions includes the water-table aquifer, streams, and recharge from precipitation. This conceptualization was necessary to simulate changes in water levels that occur in the water-table aquifer near pumping centers.

The aquifers are recharged directly in their outcrop from precipitation. The outcrop areas are shown on plates 1–5 and 7–9. Because of the subsurface truncation of the Nanjemoy and Piney Point Formations, the Piney Point–Nanjemoy aquifer does not have an outcrop area (pl. 6). Most of the water in the areas of outcrop is discharged into nearby surface-water bodies. A small amount of water percolates into the confined aquifers.

Water within the confined system moves horizontally through the aquifers and vertically through the confining units. In general, water in the downdip parts of the aquifers leaks upward through the confining units to the surficial aquifer, where it discharges to surface-water bodies.

This conceptual model is consistent with the major features of the natural flow system, as indicated on plates 1–9, as well as with the stratigraphic framework as described in table 1. However, this conceptual model is a simplification of the natural system, which is much more complex. Thus, some error in the model results may occur where the simplifications do not adequately describe the complexity of the Coastal Plain sediments.

SIMULATION OF GROUND-WATER FLOW

DATA REQUIREMENTS

In this report, the finite-difference program was employed as developed by the U.S. Geological Survey (Trescott, 1975), with modifications by Leahy (1982b). Model input requirements include (1) specifications of the hydrogeologic framework, (2) data pertaining to the hydraulic characteristics of the aquifers and confining units, (3) an initial head matrix, (4) boundary conditions, (5) space and time discretizations, and (6) locations and rates of ground-water withdrawals and recharge. The hydrogeologic framework describes the horizontal and vertical extent of aquifers and confining units, the location of their outcrops, and other pertinent features, such as areally extensive surface-water bodies. The hydraulic characteristics refer principally to the transmissivity and storage coefficient of the aquifers and to the leakance of the confining units. The initial head matrix is the head distribution for each aquifer used at the start of a model simulation. Types of boundary conditions include areas across which there is no flow or flow is constant for given periods, and areas where the head in the aquifer remains constant. Space discretization refers to the rectangular gridding both horizontally and vertically of the modeled flow system. Time is discretized into intervals referred to as "pumping periods." During each pumping period, flux into or out of the model remains constant but may be varied in successive pumping periods. Flux may include any volumetric addition or subtraction from the model, such as evapotranspiration, natural ground-water recharge, base-flow discharge, pumpage from wells, and artificial and induced recharge.

MODEL DESIGN

A principal approach used in the study was to develop a regional digital model that describes and quantifies the flow system of the entire northern Atlantic Coastal Plain sediments and that could be used, in future studies, to establish boundary conditions for areally smaller models. The first step was to calibrate a three-dimensional model that would simulate flow in a framework of 10 layers under prepumping steady-state conditions. The flow model at steady-state conditions was then converted to an 11-layer model to simulate transient conditions. This 11-layer model differs from the 10-layer model because ground-water recharge is applied, pumpage is incrementally entered for 1900–80, and discharge and recharge to the surface-water regime are represented. Therefore, not all of the assumptions and conditions used were the same for both models.

GRID

The grid of the regional model (Leahy and Martin, 1986) that simulates the flow system of the entire northern Atlantic Coastal Plain was aligned orthogonal to this flow system. The model of the Maryland, Delaware, and District of Columbia portion is a subset of the regional model. One cell of the regional model is represented by four cells in the model of this study.

The study area (pl. 1*A*) is divided into a rectangular grid of 42 rows and 36 columns, so that there are a total of 1,512 cells per layer. The rows are numbered in a southerly direction, and the columns in an easterly direction. The maximum extent of any layer, however, reduces the total number of active cells to 1,038. All cells are square and measure 3.5 mi (miles) on a side. Thus, the area that each cell represents is about 12.25 mi^2.

The aquifer system is divided into 10 layers of cells, each layer representing one modeled aquifer. The transient simulation included another set of cells that accounted for ground-water discharge derived from the unconfined aquifer. This new set of cells represented the surface-water system.

MODEL ASSUMPTIONS

A digital model is a mathematical approximation of a physical system. Such conceptual models must be amenable to mathematical analysis. The conceptual model is built on available hydrogeologic data. However, because there is a scarcity of data for much of the flow system within the study area and because the conceptual model is a simplification, many assumptions were made, as follows.

a. The head in each cell represents an average head for the aquifer in that cell. Since the area of each cell is about 12.25 mi^2, the average head is a gross approximation, especially in the western portion of the study area, where hydraulic gradients are steep.
b. Heads in the uppermost active layer represent the water table and are held constant during the simulation of prepumping conditions. These constant heads provide recharge to the confined aquifers or act as discharge cells. Typically, these cells provide about 1 in/yr or less of recharge as deep percolation to the confined aquifers.
c. During transient simulations, each water-table cell was coupled with a new set of "river-head" cells. Recharge was applied to the water-table cell. Hydraulic properties were calculated for a confining unit between the water-table and surface-water cells. Under prepumping conditions for the transient simulation, recharge from the water table to the confined flow system equaled the amount calculated by the steady-state model.
d. Hydraulic properties of the aquifers are isotropic, and all flow within the aquifers is horizontal. There is no detailed information on such properties; however, because of the model scale, this approximation is believed to be reasonable.
e. The surficial aquifer and outcrops are treated as confined aquifers; that is, transmissivity is a constant and not a function of saturated thickness. However, a large storage coefficient typical of an unconfined aquifer was specified.
f. Flow through the confining units is vertical and represents leakage between aquifers.
g. Release of water from storage and the propagation of head differences across confining units are both instantaneous.
h. The crystalline basement both underlying and along the west side of the Patuxent Formation was modeled as a no-flow boundary.
i. The locations of the 10,000-mg/L isochlors as defined by Meisler (1981, fig. 4) were modeled as no-flow boundaries. These isochlors occur at considerable depths in both layers 1 and 2.
j. In several layers, the aquifers pinch out laterally as the sand grades into silt and clay. These boundaries are modeled as no-flow boundaries.
k. The Potomac Group (Potomac Formation in Delaware) was modeled as two aquifers, equivalent to the Patuxent and Patapsco Formations where defined.
l. The Chesapeake Group was modeled as two separate aquifers equivalent to the upper and lower Chesapeake aquifers. Although the upper and lower Chesapeake aquifers are each subdivided into two or three separate aquifers locally, there is much interconnection at the scale of this model; therefore, representation of the Chesapeake as more than two layers is impractical.

INITIAL CONDITIONS

The calibrated heads of the simulation of prepumping conditions were entered as initial heads for the simulations of transient conditions. In the simulation of prepumping conditions, all of the water-table cells, both in layer 10 and in outcrop areas of the remaining 9 layers, were held constant. Thus, in the transient simulation, the initial water-table heads were identical to the input heads in the simulation of prepumping conditions.

BOUNDARY CONDITIONS

Constant-head, constant-flux, and no-flow boundaries were all used in the several phases of modeling. Figure 5 is a diagrammatic section along model row 20, extending from Baltimore on the west to the Atlantic Coast of Maryland on the east, and shows the relationships between confining units, aquifers, and various boundary conditions as conceptualized in the model.

A saltwater-freshwater interface was assumed to exist at the 10,000-mg/L isochlor as defined by Meisler (1981). This interface is extant in layers 1, 2, and 3 and, in the model, represents a no-flow boundary. No-flow boundaries were established at cells that represented the lowest active cell in any vertical section of cells. For layer 1, this cell generally represented the interface between the Patuxent Formation and the basement rock. The cells around the perimeter of the model, except for those discussed below, were constant-flux cells. The constant flux in or out of these boundary cells was calculated by the regional model and is an input to the model of this study. The procedure was an iterative process because for a number of intermediate steps, new data (for example, transmissivity and leakance) were provided from the separate subregional modeling studies. As new data were provided to the regional model, new perimeter fluxes were calculated. This process continued until the changes in perimeter flux were negligible.

Initially, the model was calibrated for prepumping steady-state conditions. Boundaries for the system were of several types. Heads at cells where water-table conditions prevailed were set to constant altitudes. All other boundary conditions were as discussed above.

To test the validity of the saltwater-freshwater interface, Leahy and Martin (1994) used a variable-density model to simulate the 1980 heads for the regional model. They found that the differences in heads between the regional model and the variable-density model generally were less than 10 ft.

For the transient simulation, the constant-head boundary cells of the prepumping simulation were altered, as indicated in figure 6, to constant-flux boundary cells. A constant flux of 13.9 ft^3/s was applied to these cells to represent that portion of precipitation that recharges the water-table cells for the transient simulation. Johnston (1977, p. 1) showed that for southern Delaware the average winter base flow of 12.6 to 14.4 ft^3/s (Johnston, 1976, p. 23; 1977, p. 7) provides a good estimate of the long-term recharge rate.

The prepumping simulation calculated the flux ("deep percolation") from the constant-head cells to the confined part of the system. In the transient model, recharge and a new cell representing surface-water flow are added. The head in this new cell is held constant, thus allowing the head in the water-table cell to fluctuate. A leakance for a confining unit between the new surface-water cell and the water-table cell is calculated as follows.

First, $$Q_b = Q_r - Q_d \tag{1}$$

where

- Q_r recharge to the water-table cell (ft^3/s),
- Q_d flux from the water-table cell to the confined part of the system (ft^3/s), and
- Q_b ground-water discharge from the water-table cell to the surface-water cell (ft^3/s).

Also, by Darcy's law,

$$Q_b = \frac{K'A(h_{wt} - h_r)}{b'} \tag{2}$$

where

- K' vertical hydraulic conductivity (ft/s),
- A area of the cell (ft^2),
- b' thickness of the confining unit (ft),
- h_{wt} head in the water-table cell (ft),
- h_r head in the surface-water cell (ft); and

$$L = \frac{K'}{b'} \tag{3}$$

where L is a leakance [(ft/s)/ft]; therefore,

$$L = \frac{Q_b}{(h_{wt} - h_r)A} \tag{4}$$

For example, in the prepumping simulation, the "deep percolation" from the water-table cell of the Patuxent aquifer in row 9, column 6, is 0.9 ft^3/s (fig. 6). In the transient simulation, 13.9 ft^3/s (15 in/yr) of recharge is applied to this cell. Subtracting 0.9 ft^3/s leaves 13.0 ft^3/s of discharge from the water-table cell to the surface-water cell of the transient model. Throughout the transient simulation, the total recharge (13.9 ft^3/s) remained constant; however, deep percolation and discharge to surface water varied with head changes.

All other boundary conditions for the simulations of transient conditions were unchanged from the prepumping simulation. The fluxes around the periphery, which were provided by the regional model, varied from one pumping period to the next. The regional model was designed so that the pumping periods were identical to those of the smaller models and, thus, provided appropriate fluxes for each of the 10 pumping periods.

MODEL CALIBRATION

Model calibration under prepumping and transient conditions was accomplished by adjusting transmissivities and leakances. The model was considered calibrated when calculated heads closely matched known heads. A

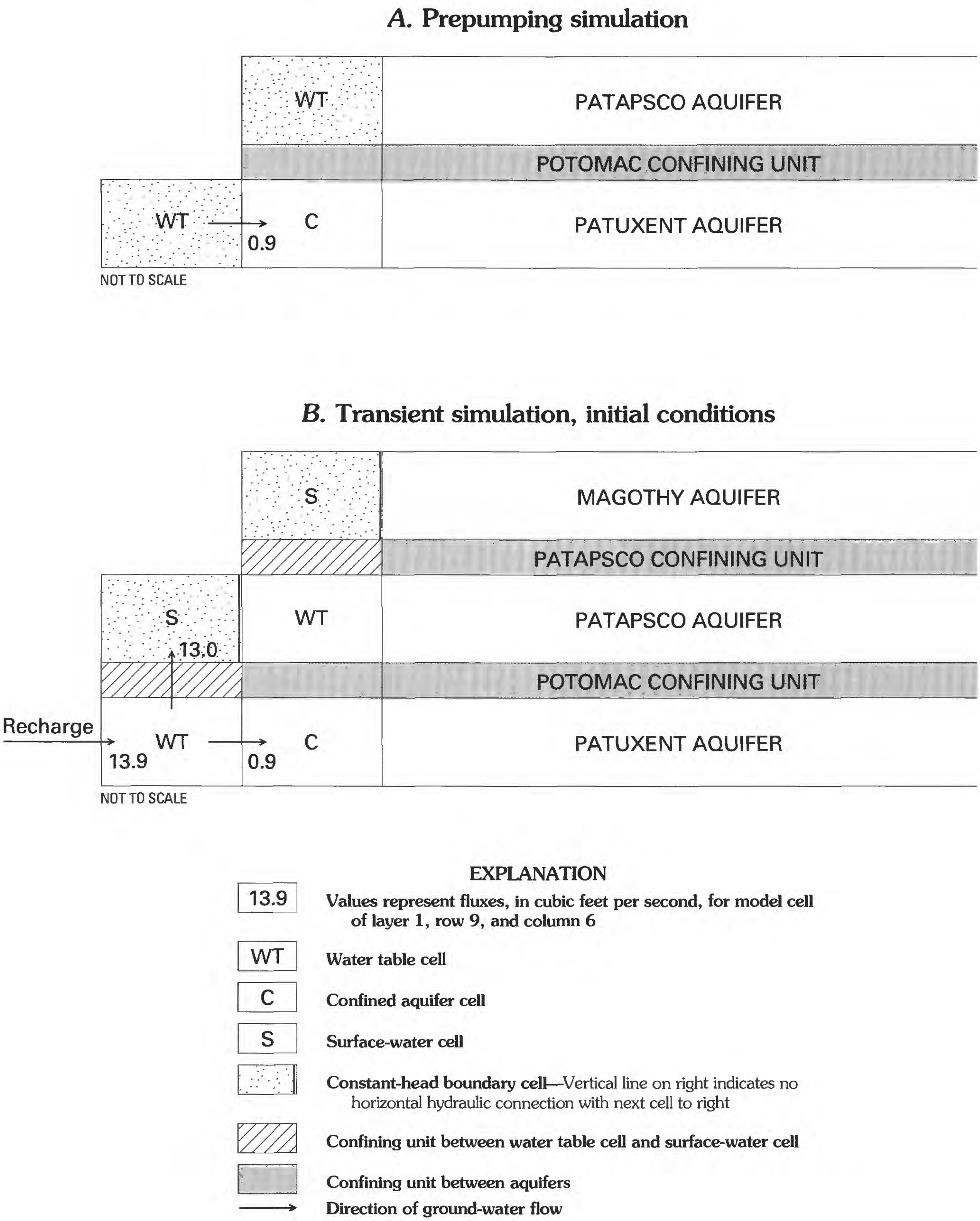


FIGURE 6.—Relation of constant-head cells for simulations of prepumping and transient conditions.

close match was defined to be 5 percent of the maximum head difference within the modeled area, or about 18 ft. Both the known and calculated heads represent an average for a given cell area of 12.25 mi^2. The 5-percent criterion is considered reasonable because heads can differ considerably (1) over an area as large as 12.25 mi^2 and (2) across the full thickness of multilayered aquifers.

PREPUMPING CONDITIONS

The model was first calibrated under steady-state prepumping conditions. To calibrate the prepumping model, either the aquifer transmissivity or the leakance of the confining units or both parameters were adjusted. The resultant head distribution was used as the initial head distribution for transient simulations. Surface-water cell leakances from the calibration for prepumping conditions were used for the transient simulation. The model was recalibrated under transient conditions and then rerun under prepumping conditions to recalculate leakance values between water-table cells and surface-water cells. This iterative process was run until a final acceptable calibrated model was achieved.

Potentiometric maps for prepumping simulations as calculated by the model are shown on plates 1–9. These maps were derived by running the model to a steady-state solution using the hydraulic properties obtained from the final transient calibration and represent an approximation of the prepumping head distribution.

Potentiometric maps of prepumping conditions were prepared during smaller scale studies for the Magothy aquifer in southern Maryland (Mack, 1974, fig. 17; Mack and others, 1981), Aquia and Piney Point aquifers in southern Maryland (Chapelle and Drummond, 1983, pls. 4, 7), the Piney Point aquifer in Maryland (Williams, 1979, fig. 7), and the Cheswold aquifer, which is part of the multilayered lower Chesapeake aquifer (Leahy, 1982a, fig. 11). Because these maps were estimated from insufficient amounts of data, they are of limited value. However, the calculated prepumping head maps shown on the plates generally show agreement with the previously published potentiometric-surface maps.

TABLE 2.—*Pumping periods for transient simulations, 1900–80*

Pumping period	Time interval	Duration, years
1	1900–20	21
2	1921–39	19
3	1940–45	6
4	1946–52	7
5	1953–57	5
6	1958–64	7
7	1965–67	3
8	1968–72	5
9	1973–77	5
10	1978–80	3

TRANSIENT PUMPING CONDITIONS

The transient simulation in which the model was calibrated represents the ground-water flow system between 1900 and 1980. Table 2 indicates the duration of each of the 10 separate pumping periods, which correspond to the pumping periods used in the regional model. For all pumping periods, hydraulic properties and recharge remained constant. However, pumpage and boundary fluxes differed.

Pumpage used for the transient simulations is given in table 3. Pumpage simulated in the model did not include nonconsumptive, shallow, small withdrawals, which are areally spread throughout the unconfined aquifers. Such pumpage generally includes domestic, small commercial, and irrigation users and has had little effect on water levels. Thus, simulated pumpage is less than the total pumpage. It is estimated, based on pumpage data reported by Wheeler and Wilde (1989), Hodges (1985), and Cushing and others (1973), that the total pumpage

TABLE 3.—*Pumpage for transient simulations, 1900–80*
[Pumpage, in cubic feet per second]

Aquifer	Pumping period									
	1	2	3	4	5	6	7	8	9	10
Surficial	0.819	4.641	7.046	9.452	12.796	14.022	17.646	20.843	28.994	39.249
Upper Chesapeake	.062	1.008	1.701	2.369	3.115	3.154	5.117	10.719	14.082	14.012
Lower Chesapeake	.152	2.856	4.476	5.360	8.129	9.935	11.382	11.481	10.069	9.551
Piney Point–Nanjemoy	.605	2.071	3.884	5.423	8.044	8.067	9.517	10.346	10.978	10.391
Aquia-Rancocas	.009	.204	1.670	2.985	4.776	5.117	5.504	6.926	8.080	8.996
Severn	.045	.310	.310	.310	.237	.211	.266	.285	.300	.368
Matawan	0	0	0	0	0	0	0	0	0	0
Magothy	.091	.899	3.925	6.787	7.259	8.639	10.418	11.961	12.464	14.254
Patapsco	.700	6.361	13.438	16.042	22.588	30.401	36.123	40.046	45.045	51.325
Patuxent	17.031	31.160	50.839	39.291	36.193	35.129	38.031	46.584	53.981	55.353
Total	19.514	49.510	87.289	88.019	103.137	115.675	134.004	159.191	183.993	203.499

for 1978–80 was about 346 ft^3/s. Thus, the simulated pumpage of 204 ft^3/s for 1978–80 (table 3) is about 60 percent of the total pumpage.

The calibration procedure of the transient simulation involved the adjustment of model properties until an acceptable match with known data was obtained. Previous simulations of the Maryland and Delaware Coastal Plain aquifers have either adjusted the vertical hydraulic conductivity of the confining layers, or both that and the aquifer transmissivity. Both Mack and Mandle (1977, p. 20) and Chapelle and Drummond (1983, p. 44) adjusted only the vertical hydraulic conductivity, suggesting that it is the least known and, therefore, the only property that should be refined. However, Martin (1984, p. 26) and Leahy (1982a, p. 34) also refined the transmissivity. Martin (1984, p. 26) found that the model was not sensitive to differences in storage coefficients and that pumpage and boundary conditions were sufficiently well known; therefore, these properties were not changed. None of these authors indicate the extent to which these properties were refined. However, Martin (1984, p. 55) suggests that the transmissivity values calculated either from aquifer-test analysis or through modeling may be in error by as much as 0.5 orders of magnitude and that leakance values may vary by as much as 1.5 orders of magnitude. In addition, Martin (1984, p. 55) indicates that the accuracy of these properties depends on the amount of data, grid spacing, and boundary conditions.

In this study, both the aquifer transmissivity and the confining-unit leakance were adjusted. In keeping with Martin (1984, p. 55), the transmissivity values were not adjusted more than 0.5 orders of magnitude. However, initial leakance values were adjusted by as much as 3 orders of magnitude. This adjustment was justified because of (1) a lack of data, (2) the variable nature of both the thickness and the vertical hydraulic conductivities of the many confining units, and (3) the large grid size. The minimum leakance values were about 10^{-14} (ft^3/s)/ft^3, which agree with minimum values reported by Mack and Mandle (1977, table 2) for the Patapsco Formation.

No other properties were adjusted. Martin (1984, p. 26) found, as did this study, that the model was not sensitive to differences in storage coefficients. Because of the relatively long pumping periods, the model essentially attained an equilibrium state; thus, calibration became insensitive to storage coefficients. Both pumpage and boundary conditions were fairly well known and therefore not adjusted.

Both the transmissivity values and the leakances for the confining units used for the final calibration in the model are shown on the plates.

Acceptance Criterion

Calibration was evaluated for the transient simulation using 121 hydrographs for observation wells located within the modeled area. Heads calculated from the transient simulation for pumping period 10, 1978–80, were plotted against the observed heads on the hydrographs. The observed and simulated heads are plotted for all 121 wells in figure 7. The criterion for calibration is that the simulated heads do not differ from the observed heads by more than 5 percent of the total prepumping head variance. This 5-percent difference amounts to 18 ft and is indicated in figure 7 by the parallel lines. About 85 percent of the simulated heads are within this 5-percent criterion.

Problems Associated with Simulated and Observed Heads

A number of values in figure 7 fall to the left of the line representing 5 percent of the total head variance. These heads were measured in the center of cones of depression where drawdowns were large. The simulated heads as calculated by the model represent an average for the entire cell area, which is 12.25 mi^2. For example, the average head in well Cd 44-14 (pl. *1H*) for 1980 is about 175 ft below sea level. The model-computed head is 67 ft below sea level. However, pumpage near the well is approximately 16 ft^3/s. The reverse situation occurs when an observation well is located some distance from the producing well. For example, observation well AA Cc 113 (pl. *1M*) is located 10,032 ft from a production well. Using a modified version of the Thiem equation (Lohman, 1972, p. 11),

$$h_p = h_a + \frac{2.3Q}{2\pi T} \log \frac{a}{4.81 r_p} \qquad (5)$$

where

- h_p calculated head at the observation well (ft),
- h_a model-computed head for the cell (ft),
- r_p distance from pumping well to observation well (ft),
- a grid spacing (ft),
- Q well discharge rate (ft^3/s), and
- T aquifer transmissivity (ft^2/d).

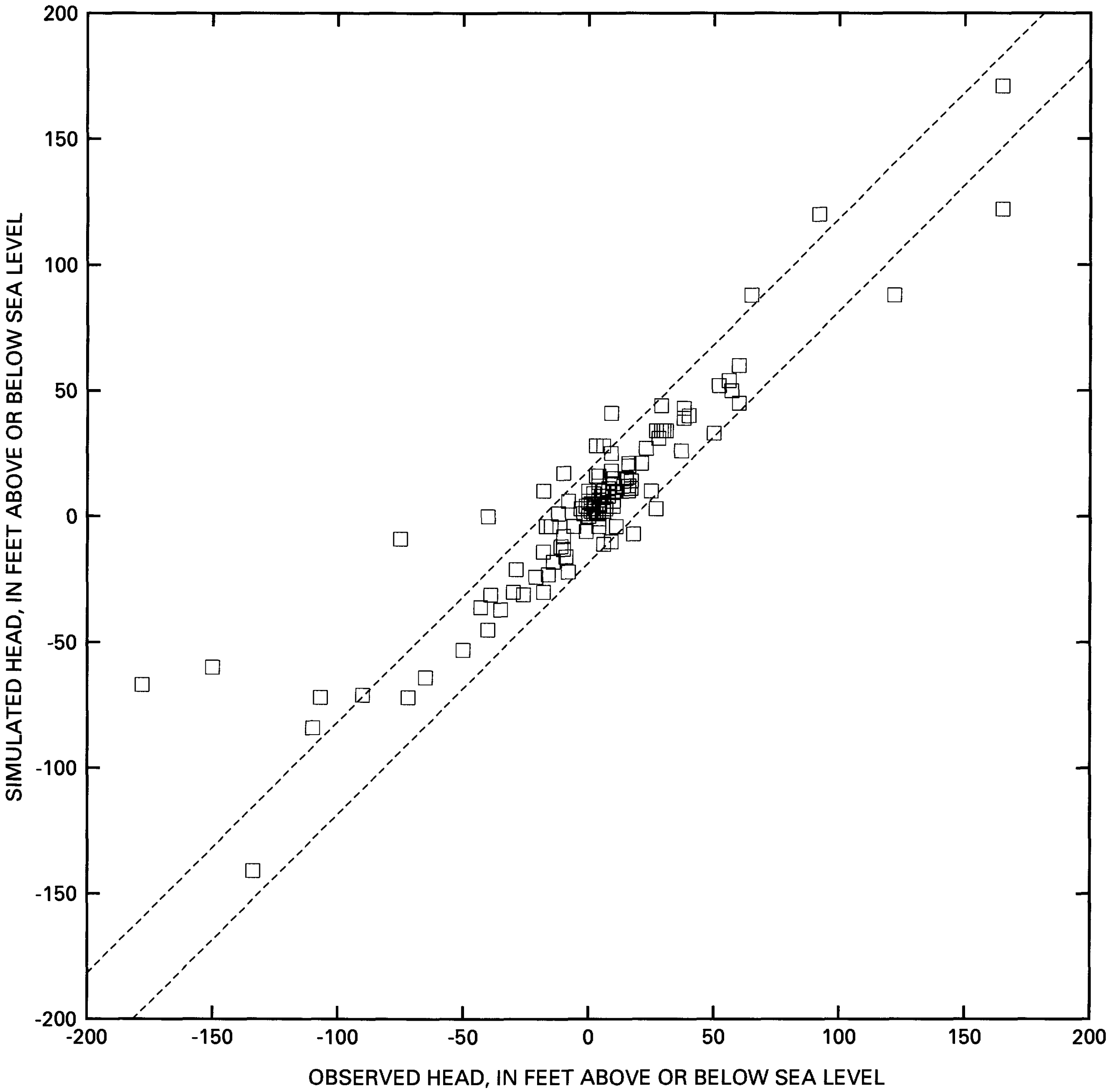
For cell row 25, column 5,

$$T = 5{,}000 \text{ ft}^2\text{/d},$$
$$Q = 1.88 \text{ ft}^3\text{/s},$$
$$r_p = 10{,}032 \text{ ft, and}$$
$$a = 18{,}744 \text{ ft;}$$

substituting gives

$$h_p = 16.33 \text{ ft.}$$

Output for the prepumping model indicates a simulated head difference under prepumping conditions of



EXPLANATION

------ Line where simulated head would differ by 18 feet from observed head

FIGURE 7.—Relation between simulated and measured heads.

about 6 ft. Adding this head difference to the simulated head under transient conditions gives h_p = 22.33 ft, which is within about 6 ft of the observed water level of 28 ft.

From equation 5, it is apparent that an observation well farther than 3,897 ft from a pumping well, both wells being located in the same model cell, would have higher measured water levels than those calculated by the model. Conversely, observation wells closer than 3,897 ft would have lower water levels. An assumption implicit in the foregoing discussion is that there is only one discharging well in the cell. Typically, because of the coarse grid spacing of the model, there may be several pumping wells within a given cell, thus reducing the effectiveness of the calibration process.

Layer 1

Plates 1*G*–*M* show hydrographs and plate 1*F* shows locations of some of the observation wells used in the calibration of the layer representing the Patuxent aquifer. The hydrographs shown on plates 1*G*, 1*I*, and 1*K* and the head map (pl. 1*N*) clearly show the result of large withdrawals in northern Delaware that have resulted in a deep cone of depression. In 1980, for example, pumpage was about 19 ft^3/s (Wheeler and Wilde, 1989). Simulated heads of the transient calibration approximately match the observed heads shown by these hydrographs.

Two observation wells (pls. 1*J* and 1*L*) for the Patuxent aquifer in southern Maryland have heads that are approximately matched by the transient pumping simulation. Observation well AA Cc 80 (pl. 1*J*) has a long-term record, and the model matches the declining heads. For well PG Ad 8 (pl. 1*L*), located in the outcrop of the Patuxent aquifer and only 35 ft deep, the model-computed heads match the long-term trend of the water table reasonably well.

The extent of calibration of the Patuxent aquifer in the model is further shown by comparing the head maps on plates 1*D* and 1*N*. However, with one exception, available data only cover the updip portion of the aquifer. A well drilled on the Eastern Shore of Maryland (Trapp and others, 1984, fig. 9) in 1981 had freshwater heads in the lower Potomac Group that increased from +15 to +53 ft with depth. The lower part of this section was saline, but heads at altitudes in the range of +15 to +35 ft corresponded to the freshwater section of the aquifer. Calibrated heads in the model were closer to the lower end of this range. The low heads in the southernmost part, plate 1*N*, are probably due to high boundary fluxes caused by pumping in Virginia and, in part, by the effects of the no-flow boundary at the saltwater interface. Plate 1*D* indicates a small cone of depression developing near Washington, D.C., evidently in response to some pumpage. However, no pumpage data were available (pl. 1*E*), and thus, the model was not calibrated in this area.

The variation of heads with depth observed in this particular well illustrates one of the problems involved in calibrating model layers, such as layers 1, 2, 8, and 9, that represent more than one aquifer. Known heads for these layers represent the heads within a given aquifer, while simulated heads from the model represent an average for the several aquifers of a given layer.

Simulated heads for any given cell represent an average for the whole cell. Measured heads in pumping wells represent the lowest heads in a cone of depression associated with such wells. Therefore, heads in or close to pumping wells will not match simulated heads for the associated cell. Well Cd 44-14 (pl. 1*H*), a case in point, was pumped at about 0.6 ft^3/s during 1978–80; consequently, the simulated heads are about 100 ft shallower than the measured heads.

Layer 2

Representative observation wells used to calibrate layer 2 are shown on plates 2*G*–*Q*, and potentiometric surface maps are shown on plates 2*D* and 2*R*. Most of these hydrographs show good agreement with the simulation results. Observation well Dc 51-9 (pl. 2*H*) matches well over the last two pumping periods, but calibration during earlier pumping periods was not possible because of a lack of historical pumping data. Wells HA Ed 24 (pl. 2*G*) and CH Dd 33 (pl. 2*P*) also typify this deficiency, namely, that pumpage data were considered to be good for the recent past but during the earlier pumping periods were often incomplete.

The two observation wells on plate 2*L* again demonstrate the difficulty of calibrating a model layer that represents several aquifers. Here, two wells located near each other are screened in two different aquifers within the Patapsco Formation. In this case, the model was calibrated to split the difference between the two records.

Comparison of the maps on plates 2*D* and 2*R* indicates the extent of agreement between the observed and simulated head distributions in the areas of northern Delaware, test well DO Ce 88, and between Washington, D.C., and Baltimore. Again, the heads observed at the well drilled on the Eastern Shore shown on plate 2*D* (Trapp and others, 1984, table 14) varied over a range of 33 ft, indicating the difficulty of developing a calibrated model for a multi-aquifer layer.

Plate 2*O* indicates the problem of coarse grid scale. The observed head in well AA Cc 40, located near the southern corner of a cell, is about 90 to 100 ft above sea level, or 30 to 40 ft below the simulated head. The simulated head in the adjacent cell to the south for the pumping period 1978–80 was simulated as 56 ft above sea level. Thus, the average head between these two cells would be about 90 ft above sea level, which matches the observed head for well AA Cc 40.

Layer 3

Layer 3, the Magothy aquifer, was not beset with the multi-aquifer problem of layers 1 and 2 because it is a single aquifer. Plates *3G–Q* are representative of the hydrographs used for calibrating the Magothy aquifer, and in general, the agreement is good between the model results and the observed heads. Large withdrawals in the Magothy aquifer have resulted in heads below sea level in significant portions of the modeled area, evident in plates *3G*, *3H*, *3K*, *3M*, *3P*, and *3Q*. Plates *3O* and *3P* show hydrographs of two observation wells located in southern Maryland, in model cell row 32, column 7. In well CH Bf 101 the simulated heads are higher than the observed heads, and in well PG Fd 41 the opposite occurs. This occurrence again illustrates that the cell scale, in which each cell encompasses an area of 12.25 mi^2, makes it nearly impossible to reproduce the exact heads.

Maps of the observed heads for 1980 and the calculated heads for the transient simulation for pumping period 10 are shown on plates *3D* and *3R*. Comparison of these two maps indicates that where observed head data are available in the area between Baltimore and Washington, D.C., simulated and observed heads match reasonably well. On the Eastern Shore, the simulation indicates a cone of depression. About 14 mi south of this cone, well DO Ce 15 (pl. *3K*) indicates a head of 15 ft below sea level, which is matched by the model. However, few data are available for the Magothy aquifer for the entire area east of Chesapeake Bay; thus, the model for this area is not calibrated.

Layer 4

Layer 4, the Matawan Formation, is generally not an aquifer in the study area but was included in the subregional model to be consistent with the regional model. Where layer 4 is an aquifer, it is used to a very limited extent and never as a major source of water supply; thus, no data were available, and the layer was impractical to calibrate.

Layer 5

Layer 5, the Severn aquifer, like the Matawan aquifer, is little used in the study area, the simulated pumpage for pumping period 10 being about 0.4 ft^3/s. One observation well (pl. *4L*) was available for use in calibration, and this hydrograph was rather well duplicated by the model. Two other data points for 1952 were available. In cell row 15, column 12, a measured head of 18 ft above sea level was simulated by the model to be 25 ft; and for cell row 19, column 9, a measured head of 4 ft above sea level was simulated to be 8 ft. Although the model showed some agreement using very few data, this layer is not considered to be calibrated.

Layer 6

Plates *5G–J* show hydrographs of four observation wells that were used for calibration of the Aquia-Rancocas aquifer (layer 6). These hydrographs indicate continuous drawdown of the Aquia-Rancocas aquifer since at least the early 1950's. Simulated heads match observed heads in the later pumping periods. The simulation for the hydrograph on plate *5J* is rather curious. Possibly, the pumpage data from 1945 to 1957 were erroneously high, resulting in the discrepancy for pumping periods 4 and 5. On the other hand, erroneously low pumpage for pumping periods 6 and 7 (1958–67) probably resulted in a mismatch for observation well KE Cd 44 (pl. *5H*).

Plates *5D* and *5K* are maps showing the observed and calculated heads for 1980. Plate *5D* indicates those areas where observed head data are available and, thus, where calibration of the Aquia-Rancocas layer was undertaken. Where data are available, the simulated potentiometric surfaces are in close agreement with measured heads.

Layer 7

Some of the hydrographs of measured heads used for the calibration of layer 7, the Piney Point–Nanjemoy aquifer, are shown on plates *6G–K*. The Piney Point–Nanjemoy aquifer has been heavily pumped near Dover, Del., near the eastern shore of the Chesapeake Bay, and in southern Maryland. Heads for all three of these areas have been measured for many years. Hydrographs on plates *6G–K* show the measured heads and the simulated heads for the transient pumping conditions. In some wells, because of the steepness of the cones of depression and because of the location of the observation wells, the simulated heads are 20 to 30 ft higher than the observed heads. Plate *6G* illustrates such a situation for well Id 55-1, which is located near the center of the cone of depression, while the simulated heads represent the average for the entire cell.

Plates *6D* and *6L* are maps of the observed and simulated heads of the Piney Point–Nanjemoy aquifer for 1980. Layer 7, because it is a single aquifer and because of a sufficiency of data, was more amenable to calibration. A comparison of plates *6D* and *6L* and the hydrographs discussed above indicates the extent of calibration of the Piney Point–Nanjemoy aquifer.

Layer 8

Both layers 8 and 9 are comprised of at least three different aquifers, and, therefore, good calibration was not tenable. However, several hydrographs for both layers indicate some degree of calibration. Plates 7*G* and 7*H* are hydrographs of observed heads in the lower Chesapeake aquifer. Plate 7*G* indicates some heavy pumping prior to 1975, but pumpage data were unavailable and therefore not simulated; thus, the calibration is only for the unstressed periods.

The two maps presented on plates 7*D* and 7*I* are potentiometric surfaces for 1977. Because of a lack of data for 1980, model results for pumping period 9, 1973–77, are shown on plate 7*I*. Plate 7*D* is based on measured heads for the Cheswold aquifer as modified from Leahy (1982a, fig. 14). The cone of depression near Dover, Del., was areally small and, because of the mesh size of the model grid, was difficult to simulate.

LAYER 9

Hydrographs are shown on plates 8*G–K*, and the head measurements for 1952 are shown on plate 8*D*. Because of a lack of data for 1980 and the availability of data for 1952, the results of pumping period 4, 1949–52, are shown on plate 8*L*. Plate 8*G* depicts two hydrographs of observed head changes—one for the Manokin aquifer and the other for the Pocomoke—and a hydrograph of the simulated heads. Plate 8*K* shows hydrographs for the Pocomoke, Manokin, and Ocean City aquifers and the head simulated by the transient model. At both of these sites, there are only small head differences between the several aquifers, and the model is in general agreement with them. A comparison of plates 8*D* and 8*L* indicates the extent of model calibration for the upper Chesapeake aquifer for this earlier pumping period.

LAYER 10

Layer 10, the surficial aquifer, occurs entirely under unconfined conditions. The initial head distribution for layer 10 that was used as input for the simulation of prepumping conditions was digitized from available water-level data and stream altitudes. For the transient simulation, about 15 in/yr of recharge from precipitation is applied to each cell of the surficial aquifer. Although in the transient simulation, about 97 percent of this recharge is discharged to surface water, the 15 in/yr is a potential source for ground-water withdrawals. Because pumpage from the surficial aquifer is very small in comparison with this recharge rate and because of high transmissivity and water-table storage, the drawdown in the surficial aquifer was negligible. The hydrographs on plates 9*G–P* demonstrate this point. The hydrographs for the surficial aquifer indicate a general agreement between the observed and the simulated heads.

SYSTEM OF REGIONAL GROUND-WATER FLOW BASED ON SIMULATIONS

PREPUMPING CONDITIONS

The final simulation of prepumping conditions used the hydraulic properties from the calibrated, transient model. The resultant head distributions for the prepumping conditions are shown on plates 1*C*, 2*C*, 3*C*, 4*E*, 4*L*, 5*C*, 6*C*, 7*C*, 8*C*, and 9*B*.

Figure 8 shows simulated areas of recharge and discharge to the confined aquifers for the prepumping conditions. In the simulation of prepumping conditions the water table was held constant. Flux into the surficial aquifer (constant-head cells) represents discharge from the confined aquifers. Although the area of individual cells is relatively large, regional patterns are clearly obvious. For example, in southern Maryland, the areas of high recharge are the uplands, and discharge is to the major rivers and the Chesapeake Bay. The pattern on the Eastern Shore is very different; low-lying areas are widely distributed in an irregular fashion, and thus, the discharge to these areas shows an irregular pattern. Recharge occurs in the higher areas between the low-lying areas and particularly along the topographic divide separating the Chesapeake and Delaware Bays.

The simulated prepumping head distribution for the Patuxent aquifer is shown on plate 1*C*. Recharge to the Patuxent aquifer is along its western outcrop. The head distribution on plate 1*C* suggests that much of the flow from the recharge areas discharged to the Chesapeake Bay. Similarly, there was a component of flow to the Potomac River. Because of the scale of the model, flow components to small rivers are not depicted. In addition, plate 1*Q* indicates a component of flow that moved downward in the aquifer under the Chesapeake Bay and the Eastern Shore. Comparison of heads in the Patuxent (pl. 1*C*) and Patapsco (pl. 2*C*) aquifers under the Eastern Shore shows that there was an upward head gradient and consequent component of upward flow through the Potomac confining unit.

Plate 2*C* shows the model-computed head distribution of the Patapsco aquifer that existed before the start of major pumping. Recharge to the aquifer occurred principally along the western boundary where the aquifer crops out. Flow within the aquifer was downgradient from the outcrop zone, and discharge occurred principally to the Chesapeake Bay, Delaware Bay, and Potomac River. There was also a major component of flow from the recharge area between Baltimore and Washington, D.C., southeastward under the Chesapeake Bay and the Eastern Shore. Along the southeastern boundary defined by the 10,000-mg/L isochlor, water in the aquifer leaked upward through confining layers into overlying aquifers.

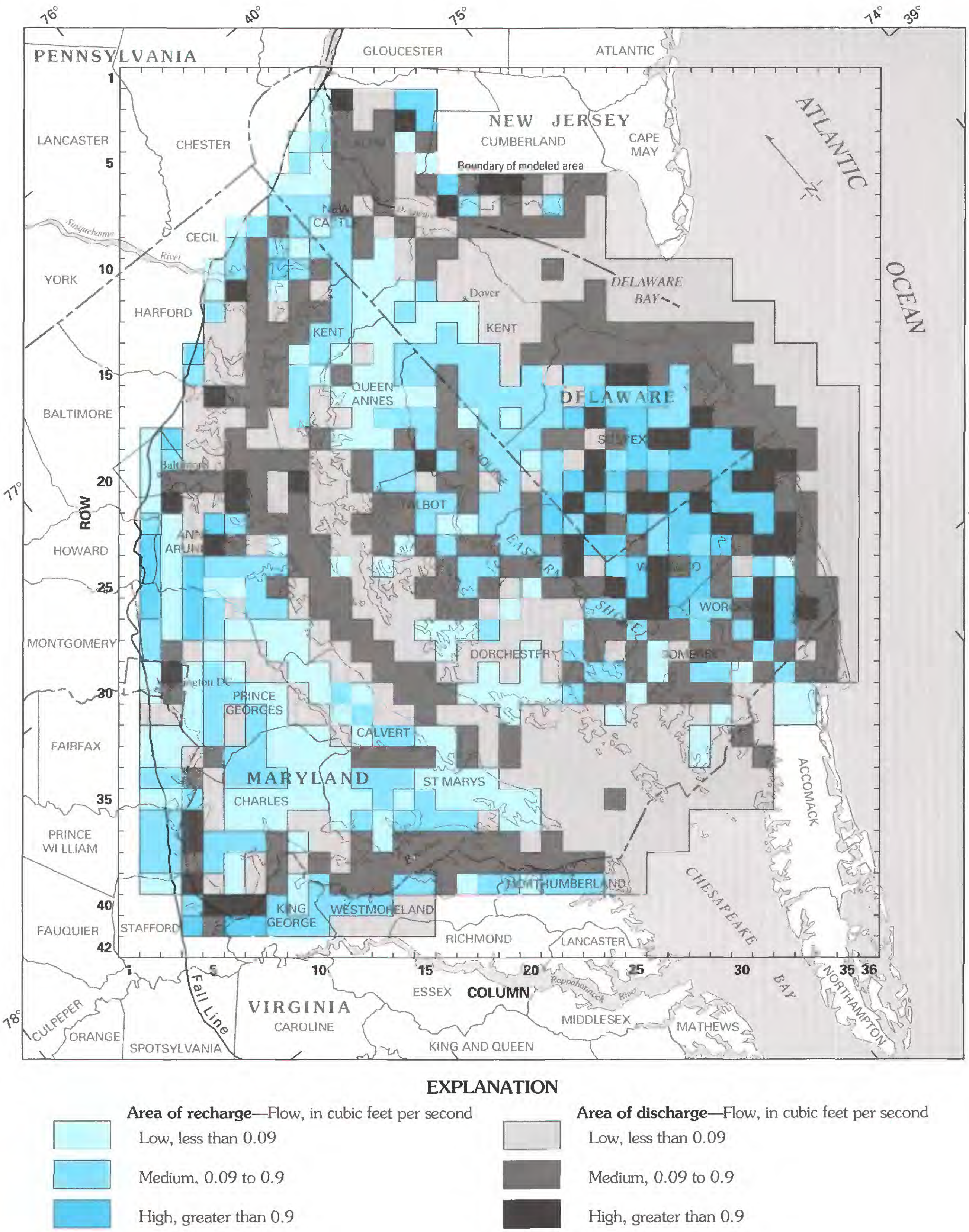


FIGURE 8.—Simulated areas of recharge and discharge to confined aquifers for prepumping conditions.

The potentiometric-surface map of the Magothy aquifer (pl. 3*C*), based on simulation of prepumping conditions, indicated a steep gradient from the west (between Baltimore and Washington, D.C.) toward the Chesapeake Bay. Recharge to the aquifer occurred in the outcrop. Water flowed downgradient to the east and the north and discharged into the Chesapeake Bay near Baltimore. On the Eastern Shore, recharge occurred along the northern part of the Maryland-Delaware line. The flow lines (pl. 3*U*) indicate that water moved both northeastward to the Delaware Bay and southwestward to the Chesapeake Bay, where it discharged. In addition, where the Magothy aquifer does not crop out, some water percolated downward into the deeper part of the aquifer. Ground water then flowed under the southern part of the Eastern Shore, where it leaked upward and discharged into overlying younger sediments.

The simulated prepumping head distribution of the Matawan aquifer is shown on plate 4*E*. Heads in the Matawan aquifer on the Eastern Shore were at an altitude as high as 25 ft, and flow in the aquifer in Delaware (pl. 4*D*) was northeasterly toward the Delaware Bay; in Maryland, flow was westerly.

Plate 4*M* shows the head distribution of the Severn aquifer under prepumping conditions. In general, flow originated in the areas where recharge to the aquifer occurred, and discharge occurred downgradient at the bays. A small amount of drawdown on the Eastern Shore of Maryland has occurred because of the pumpage.

Prepumping head distribution of the Aquia-Rancocas aquifer is shown on plate 5*C*. Flow within the Aquia-Rancocas aquifer was from the high heads associated with the recharge at the outcrop, to the discharge areas along the Potomac River to the south, to the Chesapeake Bay, and to the Delaware Bay to the north. Especially significant is the fact that the Chesapeake Bay cuts deeply into the aquifer, causing the prominent potentiometric low along the bay seen in plate 5*C*.

Plates 6*C* and 6*O* indicate the prepumping head distribution and flow paths within the Piney Point–Nanjemoy aquifer. Leakage from overlying sediments recharged updip portions of the aquifer. Flow in the aquifer was from these areas of high heads toward the Chesapeake and Delaware Bays, where water discharged either directly through the bottom sediments of the overlying bay or upward as leakage into younger formations.

The prepumping head distribution of the lower Chesapeake aquifer is shown on plate 7*C*. Recharge to the aquifer occurred within the outcrop area, and some leakage from overlying younger sediments occurred in southern Delaware and adjacent parts of Maryland. Ground-water flow in the unit was away from these areas of recharge, and discharge was to the Chesapeake and Delaware Bays and the Atlantic Ocean.

Plate 8*C* depicts the prepumping head distribution of the upper Chesapeake aquifer. The high heads were located in the area of recharge to the outcrop and where the aquifer subcrops the surficial aquifer. Flow in the aquifer was away from these high heads, discharging into the Chesapeake Bay, the Delaware Bay, and the Atlantic Ocean.

The prepumping head distribution for the surficial aquifer as shown on plate 9*B* is principally controlled by topography. Recharge to the surficial aquifer occurs throughout the area. Discharge from the aquifer is to the Chesapeake and Delaware Bays, the Potomac River, the Atlantic Ocean, and the underlying aquifers. Flow within the aquifer is from areas of high heads to low-lying discharge areas.

Simulated flux from the surficial aquifer into the confined part of the aquifer system ranged from 2×10^{-5} to 6 ft^3/s. The mean value was 0.15 ft^3/s (equivalent to 0.16 in/yr of recharge).

Table 4 indicates the total ground-water budget for all 10 layers of the model. Ground-water recharge from precipitation within the model area amounts to 11,419 ft^3/s and is the principal ground-water source (99.8 percent). A small amount, about 20 ft^3/s, enters the model through the constant-flux boundary cells from outside the area. In the model, nearly all ground water discharges to the surface-water regime. About 11,392 ft^3/s (100.0 percent) discharges as base flow for streams and rivers and as freshwater discharge to the bays and the Atlantic Ocean. A negligible amount (2 ft^3/s) is discharged out of the model through constant-flux boundary cells. Because recharge was based on winter base flow, evapotranspiration was not simulated in the model. Estimates by Johnston (1976) suggest that evapotranspiration on the Eastern Shore and in Delaware may be as much as 13 to 20 percent of the total ground-water discharge.

Ground-water flow velocities were simulated from model output. The calculations assume a porosity of 0.2. Average flow velocities ranged from 0.01 ft/d for the Matawan and Severn aquifers to 0.19 ft/d for the surficial aquifer. The maximum flow velocities were 5.3 and 4.0 ft/d in the Patapsco and Patuxent aquifers, respectively.

TRANSIENT PUMPING CONDITIONS

The prepumping conditions of the Coastal Plain were simulated as 10 separate aquifer layers. The topmost active cells, which represented the water table, were held as constant heads as a boundary condition. For the simulation of transient conditions, the water-table cells became active, and a new set of cells was added to the model (fig. 6). This new set of cells represented surface water and a new constant-head boundary. Recharge at a

TABLE 4.—*Analysis of simulated ground-water budget for prepumping conditions and pumping period 10 (1978–80)*
[—, not applicable; <, less than]

Principal sources and sinks	Prepumping				Pumping period 10 (1978–80)					
	Source		Sink		Source			Sink		
	Cubic feet per second	Percentage of sources	Cubic feet per second	Percentage of sinks	Cubic feet per second	Percentage of sources	Percentage change from prepumping	Cubic feet per second	Percentage of sinks	Percentage change from prepumping
Areal ground-water recharge........	11,419.4	99.8	—	—	11,419.4	99.6	0	—	—	—
Surface water....	0	0	11,391.5	100.0	0	0	—	11,207.8	98.0	−1.6
Constant-flux boundary.......	20.4	.2	2.3	<.1	23.2	.2	13.7	24.0	.2	943.5
Storage..........	0	0	0	0	21.6	.2	—	4.0	<.1	—
Pumpage........	—	—	0	0	—	—	—	203.5[1]	1.7	—
Total......	11,439.8	100.0	11,393.8	100.0	11,464.2	100.0	−0.1	11,439.3	100.0	.4

Percent error is 0.4 for prepumping and 0.2 for pumping period 10.
[1] Simulated pumpage represents about 60 percent of total pumpage (about 346 ft^3/s).

rate of 15 in/yr was added to all water-table cells. The prepumping model simulated the flux, or deep percolation, from the surficial aquifer cells into the confined aquifer cells. For the transient conditions, leakance values for the surface-water cells were simulated so that the difference between the 15 in. of recharge entering the model minus the deep percolation entering the confined-aquifer cells equaled the flux discharging to the surface-water bodies (see eqs. 1 to 4). The flux to the surface-water bodies is, in effect, base flow.

The average streamflow in Maryland and Delaware is about 1.0 to 1.5 $(ft^3/s)/mi^2$ (cubic feet per second per square mile) (Carpenter, 1983). It is estimated that in southern Maryland, approximately half of this flow is base flow, and on the Eastern Shore, up to about 80 percent is base flow (D.H. Carpenter, U.S. Geological Survey, oral commun., 1986; Johnston, 1971). Rasmussen and Andreasen (1959, p. 95) simulated base flow to be 0.79 $(ft^3/s)/mi^2$ for a 19.5-mi^2 drainage basin on the Eastern Shore of Maryland. Cushing and others (1973, p. 35) showed that for 10 stations for Delaware and the Eastern Shore of Maryland, base flow determined from hydrograph separation is empirically equal to the median flow. Applying this relationship to 32 stations for Delaware and the Eastern Shore (Cushing and others, 1973, table 3) indicates that base flow ranges from 0.40 to 1.21 $(ft^3/s)/mi^2$ and the weighted average, based on size of drainage area, is 0.75 $(ft^3/s)/mi^2$. However, Cushing and others (1973, p. 35) suggest that base flow is only 0.62 $(ft^3/s)/mi^2$. Base flow in the transient simulation at the end of the first pumping period for 1900–20 averaged 0.82 $(ft^3/s)/mi^2$, which seems to be a reasonable value. By the end of the transient simulation in 1980, base flow had been reduced slightly, to 0.80 $(ft^3/s)/mi^2$, a reduction equivalent to 184 ft^3/s. Simulated pumpage for the last pumping period was about 204 ft^3/s, a value that indicates that to accommodate the withdrawal by pumpage, an equivalent amount was eventually diverted from ground-water discharge to surface water. Evapotranspiration, which is estimated to be as much as 13 to 20 percent of ground-water discharge (Johnston, 1976), was not simulated and may partially account for the difference between simulated base flow of 0.80 ft^3/s and estimated values of 0.62 to 0.79 ft^3/s.

The direction of ground-water flow through confining units is controlled by head differences between the adjacent aquifers. The magnitude of flow is controlled by the leakance of the confining unit and the vertical hydraulic gradient. During the transient-conditions period of 1900–80, changes in heads because of withdrawal by pumping wells resulted in changes both in magnitude and direction of leakage through the confining units. The directions of these changes are shown on the respective plates.

Table 4 is the ground-water budget for the entire model simulation. As stated before, recharge from precipitation to the water-table cells was set at 15 in/yr. For the duration of the transient simulation, this is still the principal ground-water source and amounts to about 11,419 ft^3/s, or 99.6 percent of all sources. Withdrawal of water from storage is a minor source, totaling about 22 ft^3/s. About the same amount moves into the model through the constant-flux boundary cells.

As in the prepumping simulation, the principal ground-water sink is discharge to surface-water bodies. This sink includes discharge as base flow to streams and rivers and as freshwater discharge to the Chesapeake and Delaware Bays and the Atlantic Ocean. The total for pumping period 10 was about 11,208 ft^3/s, a reduction from the prepumping simulation of about 184 ft^3/s and a result of the pumpage of about 204 ft^3/s. In addition, a small amount of about 4 ft^3/s represents a recovery of storage as a result of some local reductions in pumpage, and some flow (24 ft^3/s) was outflow from the model area through the constant-flux boundary cells. Finally, table 4 indicates that about 204 ft^3/s was simulated as pumpage for the 1978–80 transient conditions. As noted before (see Model Calibration section), simulated pumpage is about 60 percent of the estimated actual pumpage.

In some cases, simulated drawdowns are considerably less than actual drawdowns because of both the coarseness of the mesh and the assimilation of separate water-bearing units within a single model layer. An example is the deep, but areally small, cone of depression in the Cheswold aquifer near Dover, Del. Simulated drawdowns in the lower Chesapeake aquifer, which includes the Cheswold aquifer, are less than half the actual drawdowns due to both model scale (grid coarseness) and the attempt to simulate three aquifers as a single aquifer.

Average ground-water flow velocities for the 1978–80 simulated conditions ranged from 0.19 ft/d for the surficial and Patuxent aquifers to 0.01 ft/d for the Matawan aquifer. These velocities were simulated using a porosity value of 0.2. In general, the average ground-water flow velocities were slightly greater than for the simulation of prepumping conditions.

Patuxent Aquifer

From 1900 to 1980, pumpage from the Coastal Plain aquifers has resulted in head changes in aquifers, increased flow velocities, and changes in directions of flow, both vertically through confining units and horizontally within aquifers. Plates 1*C* and 1*N* are maps of the potentiometric surfaces for the prepumping and the 1978–80 simulated results. In general, heads in the source areas have declined less than 10 ft, but where heavy pumpage occurs, several cones of depression have developed. Total pumpage for 1978–80 was about 55 ft^3/s, the majority of which occurred to the south of Baltimore and in northern Delaware (pl. 2*E*), and has resulted in heads of 100 ft below sea level.

A hydrologic-budget analysis for the last simulated pumping period (1978–80) is indicated in the table on plate 1*R*. A comparison with the prepumping budget indicates some changes, mainly due to pumpage from the Patuxent aquifer. The location of the pumpage, as shown on plate 1*E*, indicates that pumpage is entirely in the updip part of the aquifer.

The principal source of ground water in the Patuxent aquifer is recharge from precipitation in the outcrop areas. This recharge is constant throughout the simulation and amounts to about 529 ft^3/s. For pumping period 10, this recharge accounts for 97.3 percent of the sources for the pumpage from the aquifer. Increased drawdowns due to pumpage resulted in a 5.5 ft^3/s reduction in ground-water storage, which is only 1 percent of the total sources for the aquifer. Other sources are leakage through the overlying Potomac confining unit and flow through constant-flux boundary cells. As a result of pumpage, these two sources increased from about 0.2 percent of the total sources to about 1.7 percent. Plate 1*O* indicates that vertical flow throughout most of the Potomac confining unit has reversed and is now downward into the Patuxent aquifer. As a result, leakage into the Patuxent aquifer increased about eightfold from 1900 to 1980.

The principal ground-water sink for the Patuxent aquifer is discharge to surface-water bodies and amounts to about 483 ft^3/s, or 88.8 percent of the total discharge. However, this value represents a reduction of about 39 ft^3/s (8 percent) from the prepumping simulation, a result of pumpage of about 55 ft^3/s. There is a concomitant 50-percent reduction in vertical leakage out of the aquifer through the Potomac confining unit from 1900 to 1980. In some areas, a minor reduction in pumpage from pumping period 9 resulted in about 1 ft^3/s recovery of ground water in storage.

Average ground-water flow velocities for the simulated prepumping conditions were 0.12 ft/d, and for the simulated 1978–80 conditions were 0.19 ft/d, an increase of about 58 percent. This increase is due to the increased hydraulic gradients that resulted from pumpage. The most pronounced increases in flow velocities have been in northern Delaware, where velocities have increased by 2 to 3 orders of magnitude. The maximum velocity for any cell in the aquifer in the model is 4.49 ft/d. Comparing plate 1*P* with plate 1*Q* indicates some changes in flow patterns due to pumping. In plate 1*P*, three locations of heavy withdrawals located in northern Delaware, Baltimore, and south of Washington, D.C., represent the principal ground-water sinks for the pumping period 1978–80. The withdrawals have also caused the ground-water divide between Baltimore and Washington, D.C., to shift southward (pls. 1*P* and 1*Q*).

Patapsco Aquifer

A hydrologic-budget analysis for the Patapsco aquifer is presented on plate 2*V*. The prepumping budget was previously discussed. A number of notable changes from prepumping conditions to pumping period 10 have occurred, principally as a consequence of the pumpage. The location of this pumpage is shown on plate 2*E*.

Plates 2*C* and 2*R* are maps of the potentiometric surfaces for the prepumping and the 1978–80 simulations. In general, heads in the source areas have declined less than 10 ft. The total pumpage of about 51 ft^3/s is distributed throughout the updip part of the Patapsco aquifer (pl. 2*E*) and has resulted in a regional decline of heads of about 10 to 40 ft.

Ground-water sources for the Patapsco aquifer, as modeled, include recharge from precipitation, vertical leakage, flow through the constant-flux boundary cells, and withdrawal from storage. The principal source is recharge from precipitation in the outcrop areas that amounts to 794 ft^3/s. This amount of water is augmented to a small extent by vertical leakage through adjacent confining units (25 ft^3/s), withdrawal from storage (7 ft^3/s) as heads declined in response to pumpage, and constant flux (4 ft^3/s) across boundary cells. Vertical flow through the Patapsco confining unit has reversed over large areas (pl. 2*S*) and is now downward into the Patapsco aquifer. Simulated leakage into the Patapsco aquifer increased by about 30 percent from 1900 to 1980.

Ninety percent of the discharge (about 742 ft^3/s) from the Patapsco aquifer is base flow to streams and rivers and freshwater flow to the Chesapeake and Delaware Bays. However, pumpage during pumping period 10 resulted in a net reduction from prepumping conditions of about 46 ft^3/s (6 percent) of ground-water discharge to surface-water bodies. The total pumpage amounted to about 51 ft^3/s, or about 6.2 percent of the prepumping discharge from the aquifer as modeled. Vertical leakage from the aquifer through confining units was 21 ft^3/s, and flow out from the constant-flux boundaries amounted to 7 ft^3/s.

Average ground-water flow velocities (0.09 ft/d) for the last simulation period are only slightly changed from the prepumping simulation. This change is a result of an overall lowering of heads; thus, the change in hydraulic gradients was negligible. The exception is in northern Delaware, where flow velocities increased about threefold.

A comparison of plates 2*T* and 2*U* indicates some changes in ground-water flow patterns. Five cones of depression are apparent, including the aforementioned deep cone in northern Delaware and four minor cones, two of which are located in southern Maryland, one south of Baltimore, and the other on the Eastern Shore of Maryland. In addition, a greater part of the flow, especially on the Eastern Shore, is directly southward because of large ground-water withdrawals in Virginia.

Magothy Aquifer

Plates 3*C* and 3*R* are maps of the simulated potentiometric surfaces for the prepumping and the 1978–80 conditions of the Magothy aquifer. In general, heads in the outcrop areas have declined less than 10 ft; however, where heavy pumping occurs, several deep cones of depression have developed. Total pumpage for 1978–80 was about 14 ft^3/s, 20 percent of which occurred in a small area southeast of Washington, D.C., and has resulted in simulated drawdowns of 90 ft. Southeast of this area, pumpage of about 1 ft^3/s has resulted in decreases in simulated heads of about 75 ft. Southeast of Baltimore, simulated pumpage for the 1978–80 period was about 5 ft^3/s. However, because of higher transmissivities (pl. 3*A*), drawdowns have been smaller. On the Eastern Shore, pumpage of about 2 ft^3/s along the eastern side of the Chesapeake Bay has resulted in the lowering of simulated heads by up to 50 ft.

A ground-water budget analysis for the Magothy aquifer is shown on plate 3*V*. Recharge from precipitation amounting to 362 ft^3/s is constant throughout the transient simulation and, in pumping period 10, provides 95 percent of the total ground-water sources. Other sources include about 18 ft^3/s of vertical leakage through adjacent confining units, and negligible amounts from storage and constant-flux boundary cells. From 1900 to 1980, vertical leakage into the Magothy aquifer increased by about 34 percent.

Most ground water in the Magothy aquifer discharges to surface-water bodies. Pumpage of 14 ft^3/s has resulted in a 4-percent reduction of ground-water discharge to surface water. An increase from less than 0.1 ft^3/s for the simulated prepumping conditions to about 7 ft^3/s of discharge across the constant-flux boundary cells is due to large ground-water withdrawals from the aquifer in southern New Jersey and northern Virginia. Vertical leakage from the aquifer was about 11 ft^3/s.

Plates 3*T* and 3*U* reveal some interesting changes in the ground-water flow patterns. Simulation indicates that before pumping, flow in more than 50 percent of the aquifer was toward the upper reaches of the Chesapeake Bay. Plate 3*T* indicates major reversals in flow directions. Two cones of depression on the Eastern Shore are now the major sinks because almost all of the flow on the Eastern Shore and much of the flow in southern Maryland is converging to these pumping centers. The cones of depression in southern Maryland generally have greater drawdowns but affect much smaller areas. These cones of depression also have caused vertical leakage through the Matawan confining unit (pl. 3*S*) to reverse direction over extensive areas, resulting in a net increase of 5.4 ft^3/s of vertical downward flow into the Magothy aquifer. Therefore, about one-third of the pumpage comes from increased vertical flow downward through adjacent confining units.

The magnitude of ground-water flow velocities for the Magothy aquifer during the last pumping simulation is, on a regional scale, only slightly changed. Average flow velocities have increased from about 0.05 to 0.07 ft/d. Steeper hydraulic gradients near the outcrop areas, where maximum velocities for the aquifer occur, have resulted in velocity increases of about 20 to 100 percent. The maximum velocity for the period 1978–80 is about 2.9 ft/d and occurs about 15 mi south of Baltimore.

Matawan Aquifer

During the 10 simulated periods of transient conditions, the Matawan aquifer was not stressed by pumpage. However, pumpage of 3.5 ft^3/s from the overlying Piney Point–Nanjemoy aquifer (layer 7) near Dover, Del., has resulted in simulated heads in the Piney Point-Nanjemoy aquifer at altitudes as low as 90 ft below sea level. These large drawdowns in the Piney Point–Nanjemoy aquifer have resulted in head declines in the underlying Severn aquifer and, in turn, in the Matawan aquifer of up to about 30 ft. As a consequence, leakage through the confining units above and below the Matawan aquifer has increased by 30 percent (pl. 4*F*). Plate 4*F* also indicates little change in the ground-water budget from prepumping conditions to 1978–80 conditions; that is, recharge from precipitation is about equal to discharge to surface water.

Simulated ground-water flow velocities for the 1978–80 period changed little from the prepumping conditions. These velocities, averaging about 0.01 ft/d, are very low because of low hydraulic conductivities and low head gradients. However, by comparing plate 4*C* with plate 4*D*, some minor changes can be seen in the directions of ground-water flow. In particular, flow directions in the Matawan aquifer in Delaware have shifted from a northerly to a southerly direction.

Severn Aquifer

The ground-water budget (pl. 4*N*) for the Severn aquifer for the 1978–80 period changes little from the prepumping simulation. The ground-water sources include areal recharge on the outcrop, vertical leakage through the confining units, flow from the constant-flux boundary cells, and withdrawal from storage. However, 95 percent of the total ground-water budget is recharge from precipitation to the outcrop area. The principal ground-water sink for the aquifer is discharge to surface water. Discharge as vertical leakage amounts to 8 ft^3/s. The other sinks are negligible.

Average ground-water flow velocities are about 0.02 ft/d. Plate 4*I* indicates a 1978–80 discharge pattern to the streams and the bays similar to that of the prepumping simulation (pl. 4*J*). However, a number of new sinks have developed due to pumpage in the Severn aquifer or in vertically overlying aquifers. For example, in central Delaware, readjustment of heads with respect to large head changes in the overlying Piney Point–Nanjemoy aquifer has resulted in a ground-water sink in the Severn aquifer. This sink has caused changes in flow directions over a sizeable part of Delaware. Three new sinks have developed on the Eastern Shore of Maryland, the northernmost due to pumpage of about 0.3 ft^3/s from the aquifer. South of this small cone of depression is a cone caused by drawdowns in both the overlying Aquia-Rancocas aquifer and underlying Magothy aquifer, and farther south is another sink caused by head changes in the Magothy aquifer.

Aquia-Rancocas Aquifer

A comparison of plates 5*C* and 5*K* indicates that along the western margin of the Aquia-Rancocas aquifer, simulated heads have only declined slightly from the prepumping conditions through 1978–80. However, in the downdip part of the aquifer, head declines have been substantial. Simulated pumpage of about 5 ft^3/s in southern Maryland has resulted in drawdowns of about 100 ft. On the Eastern Shore, simulated heads in the aquifer have declined in response to simulated pumpage in both the Aquia-Rancocas and the overlying Piney Point-Nanjemoy aquifer.

Plate 5*E* indicates that pumpage in the Aquia-Rancocas aquifer was more areally extensive than in any of the other modeled aquifers. Although total simulated pumpage for the 1978–80 period was only 9 ft^3/s (pl. 5*O*), it was distributed over 77 model cells. Minor changes in the ground-water budget that account for this pumpage are indicated on plate 5*O* and principally include diversion of base flow and a slight change in the flow through confining units.

The ground-water budget (pl. 5*O*) also indicates that the principal ground-water source is areal recharge to the outcrops. The simulated areal recharge amounts to about 641 ft^3/s and accounts for 95.2 percent of the ground-water sources for the Aquia-Rancocas aquifer. Head changes during the 1978–80 pumping period resulted in a minor amount of ground water released from storage. Plate 5*L* indicates that much of the vertical flow through the overlying Nanjemoy-Marlboro confining unit has reversed to a downward direction. However, vertical leakage through confining units into the aquifer is slightly reduced from the prepumping simulation, probably because of slightly greater pumpage in the Piney Point–Nanjemoy and Magothy aquifers.

Ground-water discharge from the Aquia-Rancocas aquifer includes discharge to surface-water bodies, vertical leakage through confining units, and pumpage. As a direct result of 1978–80 simulated pumpage (9 ft^3/s), ground-water discharge to surface water is reduced from the simulated prepumping condition by about 6 ft^3/s (1 percent). This discharge to surface-water bodies amounts to about 632 ft^3/s and accounts for 93.7 percent of the total ground-water discharge from the aquifer.

Simulated vertical leakage out of the aquifer was reduced by 0.6 ft^3/s, from 34.0 ft^3/s to 33.4 ft^3/s.

Simulated ground-water flow velocities for the 1978–80 period are approximately the same as for the prepumping conditions. Average and maximum flow velocities for the simulated pumping period 1978–80 are 0.04 and 0.35 ft/d, assuming a porosity of 0.2. In the vicinity of the cone of depression located in southern Maryland, flow velocities have increased from 0.01 to about 0.16 ft/d. The higher velocity is equivalent to about 60 ft/yr.

As discussed previously, simulation suggests that ground water had discharged from the Aquia-Rancocas aquifer to streams in the outcrop areas, the Potomac River, and the Chesapeake and Delaware Bays before pumping. Plate 5*N* indicates that the area of the aquifer underlying the Chesapeake Bay and the Potomac River was a well-defined sink. Here, ground water was discharged either directly into these surface-water bodies or as upward leakage through confining silt and clay.

Comparing 1978–80 conditions (pl. 5*M*) with the prepumping conditions (pl. 5*N*) shows that the pattern of simulated ground-water flow during pumping conditions has drastically altered. Only at the most headward reaches of the Chesapeake Bay and the Potomac River does ground water discharge. Elsewhere, ground water has ceased to discharge to and flows under these water bodies, and now discharges to one of some half-dozen cones of depression (pl. 5*N*). It is interesting to note that flow under the Chesapeake Bay is eastward in the upper reach, while it is in the opposite direction under the lower reach.

Piney Point–Nanjemoy Aquifer

Total simulated pumpage from the Piney Point–Nanjemoy aquifer for pumping period 10 (pl. 6*P*) was about 10 ft^3/s. Although this value is slightly more than that for the Aquia-Rancocas aquifer, pumpage in the Piney Point–Nanjemoy aquifer is more concentrated, as indicated on plate 6*L*. Comparing the prepumping heads with the 1978–80 simulated heads (pls. 6*C* and 6*L*) reveals that the aquifer has undergone substantial head declines. Three distinct coalescing cones of depression have developed. To the north, centered in the Dover, Del., area, pumpage of about 3.5 ft^3/s has resulted in simulated drawdowns of about 130 ft (pl. 6*G*). Southwestward, near the eastern shore of the Chesapeake Bay, pumpage of about 3.0 ft^3/s has caused another deep cone. A third, less distinct cone has developed in southern Maryland, partly because of pumpage of about 0.8 ft^3/s and partly because of head adjustments caused by large head declines in the underlying Aquia-Rancocas aquifer. Between all three of these pumping centers, heads are below sea level.

Unlike the other confined aquifers, the Piney Point–Nanjemoy aquifer does not crop out; thus, there is no direct recharge from precipitation to the aquifer (pl. 6*P*). All ground water enters the aquifer as leakage through confining units. In the prepumping simulation, the leakage in was closely balanced by leakage out: 90.1 and 97.5 percent, respectively. However, pumping withdrawals of about 10 ft^3/s have caused a concomitant net change in leakage through confining units of about 9 ft^3/s. Areas where the direction of vertical leakage through the lower Chesapeake confining unit has reversed are shown on plate 6*M*. Water derived from storage during the last simulation period, 1978–80, was negligible.

Average ground-water flow velocities for the Piney Point–Nanjemoy aquifer for the simulation period 1978–80 are 0.14 ft/d, about 50 percent greater than for prepumping conditions. It should be noted that the model scale (that is, large-scale area) is such that the real flow velocities in the vicinity of pumping wells are orders of magnitude higher than simulated flow velocities. Simulated flow velocities on the Eastern Shore and in Delaware were an order of magnitude greater for the 1978–80 period as compared with the prepumping period. However, an average flow velocity of 0.14 ft/d (or less if porosities are greater than 0.2) is still rather low. At this velocity, it would take a particle of water more than 100 years to travel only 1 mi.

Comparison of plates 6*N* and 6*O* reveals some interesting changes in flow patterns. The prepumping simulation shown on plate 6*O*, as discussed before, indicates a flow pattern in which ground water is discharging through confining units to the Chesapeake and Delaware Bays and major rivers. In plate 6*N*, it is evident that ground-water flow under pumping conditions is now generally to the three major pumping centers. In addition, two other sinks caused by pumpage are apparent.

Lower Chesapeake Aquifer

Plates 7*C* and 7*I* are maps of the potentiometric surfaces for the prepumping and 1973–77 simulated conditions. The lower Chesapeake aquifer, model layer 8, includes three water-bearing units: the Cheswold, Federalsburg, and Frederica aquifers (table 3). Thus, simulated heads represent an approximate average of the heads in the three aquifers. For example, in the Cheswold aquifer near Dover, Del., a cone of depression existed with heads of about 75 ft below sea level (Cushing and others, 1973, pl. 6). However, at the same location, heads in the Federalsburg and Frederica aquifers were about 20 and 25 ft above sea level, respectively (Cushing and others, 1973, pls. 7–8). Simulated heads (pl. 7*I*) were about 10 ft below sea level. Near Cambridge, Md., in 1970, heads in the Cheswold and Federalsburg aquifers were about 25 ft below sea level, and in the Frederica aquifer, at about sea level (Cushing and others, 1973, pls. 6–8) because of pumpage in the Piney Point aquifer. However, from 1970 to 1980, decreased pumpage in the Piney Point aquifer from 4.4 to 1.7 Mgal/d (Wheeler and Wilde, 1989) resulted in simulated head recoveries of about 50 ft in the lower Chesapeake aquifer. Simulated heads for the lower Chesapeake aquifer for 1973–77 (pl. *7I*) were about 5 ft above sea level.

The ground-water budget based on simulations for the 1978–80 pumping period is indicated on plate *7M*. The principal ground-water source, areal recharge from the aquifer outcrop, accounts for 83.6 percent of the total ground-water source for pumping period 10 and amounts to 251 ft^3/s. About 43 ft^3/s of vertical leakage through adjacent confining units and 6 ft^3/s through the constant-flux boundary cells account for the rest of the sources for the lower Chesapeake aquifer.

Ground-water discharge from the lower Chesapeake aquifer includes discharge to surface-water bodies overlying the aquifer outcrop, vertical flow through confining units, flow through constant-flux boundaries, and withdrawal from wells. The largest single component of the ground-water sinks under pumping conditions is still discharge to streams, the Chesapeake and Delaware Bays, and the Atlantic Ocean. This component amounts to about 259 ft^3/s and represents 86.5 percent of the total, a reduction of only 2 ft^3/s from the prepumping simulation, although withdrawal because of pumpage was about 10 ft^3/s. Leakage from the aquifer as vertical flow through the confining units was about 29 ft^3/s and represents a net reduction of about 9 ft^3/s from the prepumping simulation. Areas where reversals in the direction of vertical leakage through the overlying St. Marys confining unit have occurred are shown on plate *7J*. Less than 1 percent of the discharge from the aquifer is across the constant-flux boundary.

Simulated ground-water flow velocities for 1978–80 did not change from the prepumping conditions. The average flow velocity for the 1978–80 simulation period was about 0.05 ft/d for a porosity of 0.2, or 1 mi in 290 years.

Plates *7K* and *7L* indicate ground-water flow patterns for the prepumping and 1978–80 simulations. The most obvious change is in central Delaware. At model cell row 12, column 17, in the vicinity of Dover, pumpage of 4.4 ft^3/s has caused a simulated drawdown of 37 ft (pl. *7K*). Actual drawdown in the Cheswold aquifer was about 95 ft (Leahy, 1982a, figs. 11, 14) and about zero feet in the Frederica aquifer (Cushing and others, 1973, pl. 8). Plate *7L* indicates that ground-water flow in this area was eastward before pumping, and discharged to the Delaware Bay. For the last simulation period, 1978–80, ground-water flow throughout most of central Delaware is diverted toward the pumpage center in the Dover area, as indicated on plate *7K*.

Upper Chesapeake Aquifer

Plate *8P* indicates that pumpage from the upper Chesapeake aquifer for the model simulation of 1978–80 conditions was 14 ft^3/s. The distribution of this pumpage is shown on plate *8E*. About 60 percent (8.6 ft^3/s) of the pumpage from the aquifer is along the Atlantic Coast. Because of the discontinuous nature of the upper Chesapeake confining unit separating the upper Chesapeake and surficial aquifers, vertical flow through this confining unit is a major factor affecting the heads in the upper Chesapeake aquifer. Simulated head changes because of pumpage for the 1978–80 pumping period have been rather slight. The largest simulated drawdowns, which are along the Maryland Coast, have not exceeded 10 ft.

An analysis of the ground-water budget is given on plate *8P*. Ground-water sources include areal recharge from precipitation on the outcrops, vertical flow through adjacent confining units, and flow through constant-flux cells. Simulated recharge is about 209 ft^3/s for the upper Chesapeake aquifer and represents 62.1 percent of all ground-water sources for the aquifer. As mentioned above, the upper Chesapeake confining unit is both leaky and discontinuous; thus, leakage as vertical flow into the aquifer represents 36.1 percent of the total ground-water sources and amounts to about 122 ft^3/s. Reversals in the direction of vertical flow through the upper Chesapeake confining unit (pl. *8M*) have only been downward. As a result, there has been a net increase of vertical leakage into the upper Chesapeake aquifer of 9.3 ft^3/s from the prepumping to the 1978–80 pumping conditions. A small amount of recharge occurs through the constant-flux boundary cells and accounts for less than 2 percent of the total ground-water sources for the aquifer.

Simulated pumpage of 14 ft^3/s accounts for 4.2 percent of the ground-water discharge from the upper Chesapeake aquifer. This pumpage has resulted in some decrease of discharge to surface-water bodies and a net increase of vertical flow through confining units to the aquifer. Ground-water flow to surface-water bodies decreased from about 212 ft^3/s (prepumping conditions) to about 209 ft^3/s for the 1978–80 pumping period. Simulated vertical flow through confining units into and out of the aquifer was less for 1978–80 than for the prepumping conditions. However, the net change was an increase of about 13 ft^3/s of vertical flow into the aquifer. Because head changes were minimal, change in storage was also minimal and accounts for less than 0.1 percent of the ground-water budget.

Average ground-water flow velocities for the upper Chesapeake aquifer are about 0.13 ft/d. Plate *8N* shows the flow pattern for the 1978–80 simulation. Comparing this figure with plate *8O* reveals several minor changes, the most notable being along the Atlantic Coast of

Maryland, where a reversal from seaward to landward ground-water flow has occurred.

Surficial Aquifer

Plate 9*Q* is the ground-water budget as simulated by the model for the surficial aquifer. Recharge to the aquifer by precipitation of 15 in/yr amounted to 8,384 ft^3/s and accounted for 98.6 percent of the total ground-water sources for the aquifer. Vertical flow of 112 ft^3/s through underlying confining units into the aquifer accounted for only 1.3 percent of the total sources. Change in storage was negligible and accounts for less than 0.1 percent of the budget.

The budget on plate 9*Q* indicates that the principal ground-water sink is the discharge of about 8,283 ft^3/s to surface water. This value represents a reduction from the simulated prepumping rate of 69 ft^3/s, although it is a reduction of less than 1 percent. Ground-water evapotranspiration, which is a major sink, was not simulated. Total pumpage from the surficial aquifer for 1980 is estimated to be about 122 ft^3/s (Wheeler and Wilde, 1989; Hodges, 1985, p. 167; Cushing and others, 1973, p. 5). In the model, only the major industrial and public-supply withdrawals were simulated. Simulated pumpage from the surficial aquifer for 1978–80 was about 39 ft^3/s, which represents about 19 percent of the total simulated pumpage for all aquifers, but only 0.5 percent of the total ground-water discharge from the surficial aquifer. As a consequence, the head changes in the aquifer were minimal; thus, depletion of ground-water storage was less than 0.1 percent. Vertical leakage out of the aquifer through underlying confining units was about 162 ft^3/s. The net change of vertical leakage out of the aquifer was an increase of about 40 ft^3/s, a result of pumpage in some of the underlying aquifers.

For the 1978–80 simulation period, maximum and average flow velocities were the same as for the simulation of prepumping conditions, namely, 1.58 and 0.19 ft/d, respectively. Comparison of the two velocity maps, plates 9*D* and 9*E*, shows only a few minor changes in direction and magnitude of flow.

SENSITIVITY ANALYSIS

A sensitivity analysis was performed on the ground-water flow model to determine the response of the model, as represented by changes in simulated head, to changes in model inputs. The model for transient conditions was tested for sensitivity to (1) hydraulic characteristics, by changing both the transmissivity and leakance inputs; (2) the magnitude of pumpage withdrawals; and (3) a change in boundary fluxes. Estimates of transmissivity are expected to be in error by less than 0.5 orders of magnitude, and those of leakance between aquifers to be within 1 to 1.5 orders of magnitude of the actual values (Martin, 1984, p. 55). The accuracy of the final calibrated estimates of these values will depend on the grid spacing and boundary conditions of the model and the amount and accuracy of the original data. Model calculations of base flow indicate that the simulated confining-unit leakance between the surface-water cells and water-table cells is probably in error by less than 0.5 orders of magnitude. It is estimated that domestic use of ground water in the Coastal Plain may be as much as 30 percent of the total pumpage (Wheeler and Wilde, 1989). Domestic pumpage, because of the nature of its areal distribution, probably has only a negligible effect on water levels; thus, for the sensitivity test, pumpage was varied only in areas of known large pumpage. The constant-flux boundary conditions were obtained from the regional model. Because of the iterative process used to determine these fluxes, it was assumed that the specification of these fluxes is in error by less than 0.5 orders of magnitude. The ground-water budgets (see pls. 1–9, and table 4) indicate that water derived from storage is generally less than 0.1 percent of the total budget; thus, the sensitivity of storage coefficients was not tested.

To test the effects of these possible ranges in values, transmissivity was both increased and decreased by 0.5 orders of magnitude for all model cells. The results of the sensitivity tests are shown in figure 9 for two cells in the Patuxent aquifer, one close to (fig. 9*A*) and the other far from (fig. 9*B*) any pumpage. The cell represented in figure 9*A* is located next to a cell where pumpage for pumping period 10 was 5.2 ft^3/s. To test the effects of leakance, the leakance of the confining units between aquifers for all model cells was both increased and decreased by an order of magnitude. The same was done for the leakance of the confining unit between the surface-water and water-table cells. To test the effects of pumpage on the model, the pumpage from 1900 to 1980 was first increased and then decreased by 20 percent. The effects of boundary flux specified by the results of the regional model were tested by both increasing and decreasing the flux by an order of magnitude. The results shown in figure 9 are typical of all model layers.

The saltwater-freshwater interface was represented as a no-flow boundary. The sensitivity of this boundary was tested for the regional model (Leahy and Martin, 1994), and results indicated that the sensitivity to location of the seaward boundary was minimal except for cones of depression located near the interface. The most affected cones were located in deeper aquifers and had simulated heads that were as much as 25 ft different from the calibrated heads.

TRANSMISSIVITY

As expected, the sensitivity of simulated heads to changes in transmissivity is great, especially where there are large withdrawals and associated drawdowns. To test the sensitivity of the model, transmissivity for all aquifers was both (1) reduced to one-fifth the calibrated value and (2) increased fivefold. When the aquifer transmissivity is one-fifth of the calibrated transmissivity values, the simulated drawdown is considerably greater. For example, figure 9*A* indicates that as pumpage is increased to 5.2 ft^3/s (pumping period 10) and transmissivity is reduced, model heads are reduced an additional 165 ft. Even in areas where pumpage is negligible, reducing transmissivity will result in a changed head distribution. Figure 9*B* shows that reduced transmissivity results in the lowering of heads by about 35 ft. Increasing transmissivity has the opposite effect but is not nearly as pronounced.

LEAKANCE

The effects of change in leakance values are somewhat more complicated. The leakance values simulated for the surface-water cells control the amount of recharge or deep percolation to the confined aquifers. Recharge of 15 in/yr is applied to all water-table cells. Most of the recharge passes through the confining layer to the surface-water cells, and in general, less than 1 in/yr goes into the confined aquifers. The result of an order-of-magnitude reduction for all surface-water cells is a considerable reduction in flux to the surface-water bodies, and as a consequence, a large amount of water is added to the confined part of the aquifer system. The obvious result of this additional recharge is much higher heads. The increased-head distribution is propagated throughout the model, as shown in figure 9. Similarly, an order-of-magnitude increase in leakance values results in the lowering of heads throughout the model. Figure 9*A* indicates that pumpage has no effect on the amount that heads are increased or decreased.

Leakance of the confining units between aquifers was tested by both increasing and decreasing the leakance values for all confining units by an order of magnitude. The results are shown in figure 9. Increasing the leakance values resulted in increased heads of as much as 65 ft, while decreasing leakance values resulted in decreased heads of about 4 ft for pumping period 10.

Walton (1970, p. 364) lists recharge rates from different studies. These values range from 0.4 to 5.4 in/yr, which is greater than the 0.15 in/yr of deep recharge in this model. This discrepancy would indicate that the deep percolation, as calculated from the simulation of prepumping conditions, may be low. However, Walton's recharge rates may include recharge to shallow or intermediate flow regimes. Both Chapelle (U.S. Geological Survey, oral commun., 1984) and Achmad (Maryland Geological Survey, oral commun., 1984), in their model studies of the Maryland Coastal Plain, found that the deep percolation in their ground-water flow models was consistent with that used in this study. It is apparent, however, that the model is very sensitive to this deep percolation and that higher recharge rates would mean that the transmissivities used in this model would probably be low.

An increase in leakance values has less effect on heads in the model. Because most of the 15 in/yr of recharge is already discharged to the surface-water cells, an increase in leakance does not materially change the amount of deep percolation. Figure 9 indicates that the heads are reduced about 10 to 15 ft, in contrast to an increase of 50 to 90 ft when the leakance is reduced.

PUMPAGE

Total simulated pumpage was 203.5 ft^3/s. The result of both increasing and decreasing pumpage by 20 percent, as shown in figures 9*A* and 9*B*, was to change the head for pumping period 10 by 15 and 7 ft, respectively. Increased pumpage resulted in increased drawdowns, and correspondingly, decreased pumpage resulted in decreased drawdowns.

BOUNDARY FLUX

The flux along the model boundary was simulated by the regional model and specified for each pumping period. As indicated earlier, these regionally derived fluxes are probably in error by less than 0.5 orders of magnitude. In the sensitivity analyses, the fluxes for each pumping period were changed by an order of magnitude, and the results are shown on figure 9. The effect of increasing or decreasing these fluxes differs; the changes in head can be as much as 200 ft and as little as 2 ft, depending on the location of pumping stresses and hydrologic characteristics of aquifers.

In summary, the model is most sensitive to (1) a decrease in transmissivity, (2) an increase in transmissivity, (3) a decrease in leakance of the surface-water cells, (4) an increase in leakance of the confining units between aquifers, and (5) an increase in specified boundary fluxes. The model as is understates the amount of pumpage and possibly the rates of deep percolation. However, these two factors have countereffects and tend to balance each other. The hydraulic properties used in the model probably represent a reasonable approximation of the flow system.

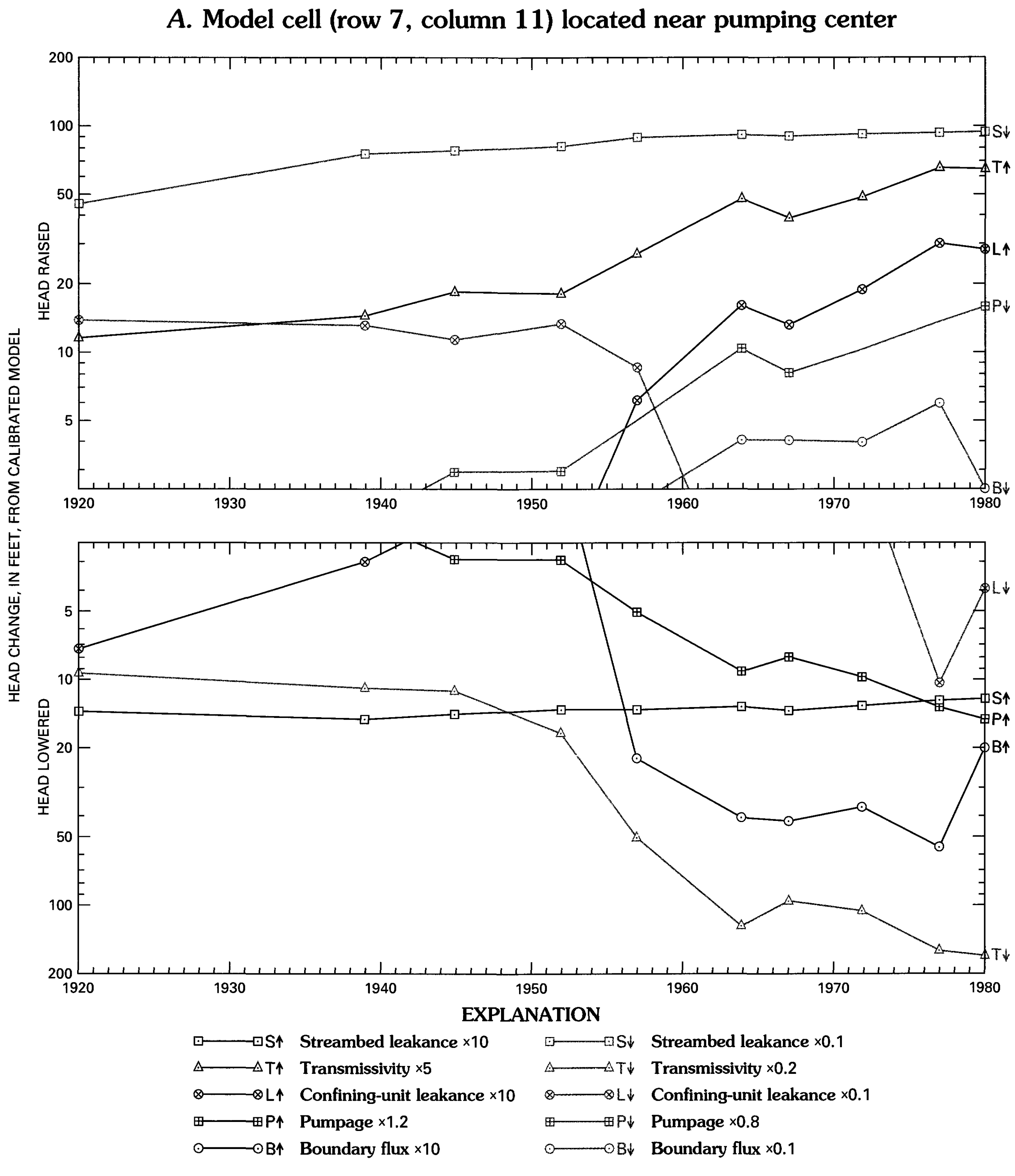


FIGURE 9.—Analyses of sensitivity to changes in transmissivity, leakance, pumpage, and boundary fluxes at two cells in the Patuxent aquifer: *A*. Model cell (row 7, column 11) located near pumpage.

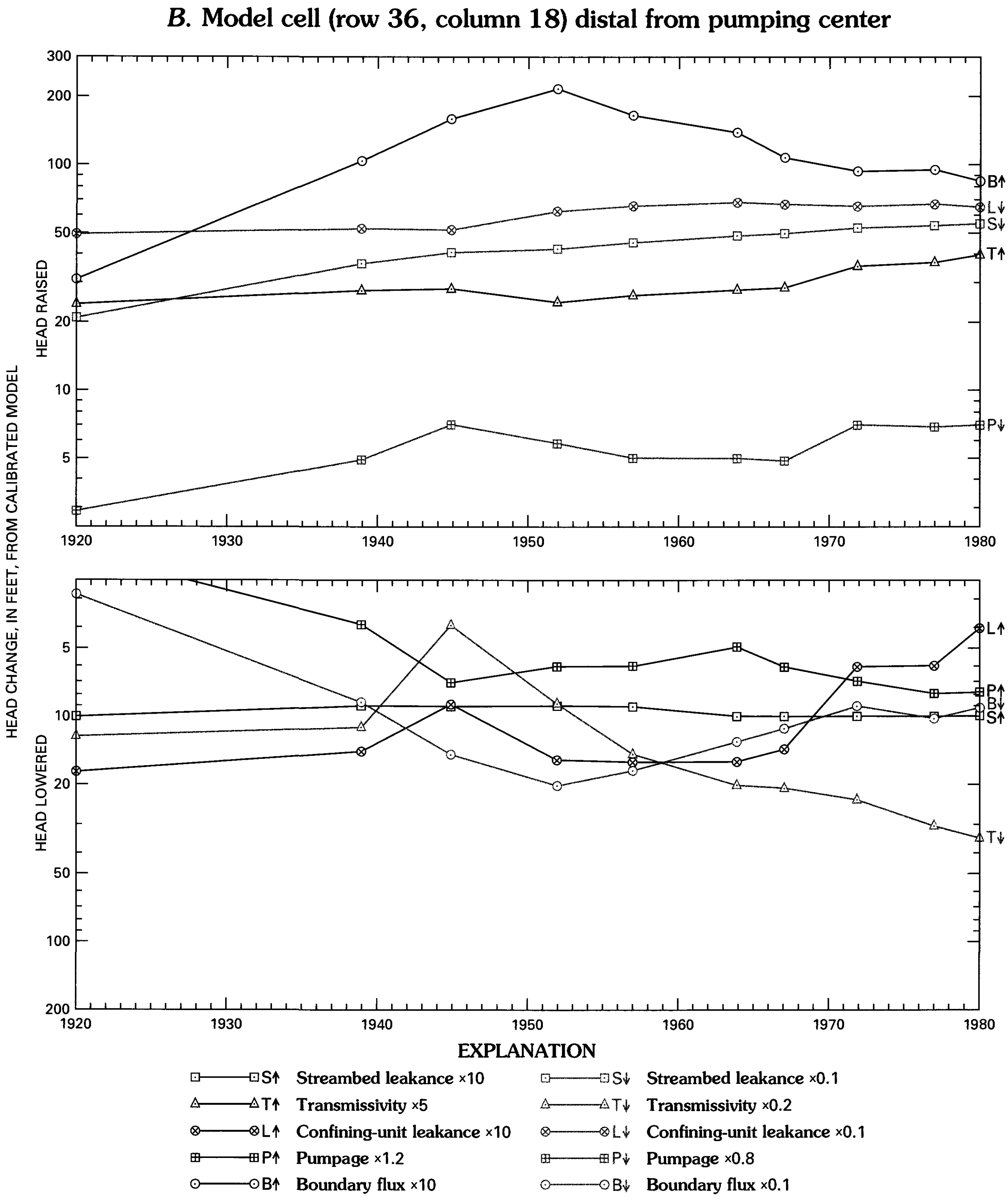


FIGURE 9.—*B*. Model cell (row 36, column 18) distal from pumpage.

SUMMARY AND CONCLUSIONS

The Coastal Plain sediments of Maryland, Delaware, and the District of Columbia consist of a complex sequence of unconsolidated sand, silt, clay, and gravel of Early Cretaceous to Quaternary age, ranging in thickness from a featheredge at the Fall Line in the northern and western parts of the study area to about 8,500 ft in the southern and eastern parts. The sediments vary from marine to fluvial in origin and generally are underlain by Precambrian crystalline basement rocks. The coarser grained layers of the sequence act as aquifers, capable of providing significant amounts of water, whereas the finer grained layers act as confining units separating the aquifers. Because the entire sequence dips at a low angle to the south and east, the aquifers crop out in broad bands roughly parallel to the Fall Line, with the oldest aquifers cropping out nearest the Fall Line.

The 10 major aquifers delineated during this study are, from oldest to youngest, the Patuxent, Patapsco, Magothy, Matawan, Severn, Aquia-Rancocas, Piney Point–Nanjemoy, lower Chesapeake, upper Chesapeake, and surficial aquifers. The areal extent of each aquifer differs, ranging from 1,071 mi^2 for the Matawan aquifer to 10,775 mi^2 for the Patapsco aquifer. Most of the aquifers pinch out at the edge of their outcrop areas and thicken to the south and east; the Patapsco aquifer reaches the greatest thickness, about 1,700 ft. The aquifers are separated by nine major confining units, which are, beginning with the oldest (which overlies the Patuxent aquifer), the Potomac, Patapsco, Matawan, Severn, lower Brightseat, Nanjemoy-Marlboro, lower Chesapeake, St. Marys, and upper Chesapeake confining units. The confining units, like the aquifers, are of different areal extent and thickness, the thickest being the Potomac (700 ft).

The hydraulic properties of the sediments differ both within and between each aquifer and confining unit. The maximum transmissivity used in the ground-water flow model constructed during this study is in the Patapsco aquifer (23,200 ft^2/d), followed by the surficial aquifer (19,700 ft^2/d) and the Patuxent aquifer (13,900 ft^2/d). Locally, higher transmissivities have been documented by field tests. The leakance of the confining units is lowest [10^{-14} (ft^3/s)/ft^3] in the Potomac, Nanjemoy-Marlboro, and Chesapeake confining units and highest [10^{-3} (ft^3/s)/ft^3] in the Patapsco, Matawan, lower Brightseat, Nanjemoy-Marlboro, and St. Marys confining units. Within aquifers, transmissivity can differ by as much as 2 orders of magnitude; within confining units, leakance can differ by as much as 11 orders of magnitude.

Large quantities of ground water have been withdrawn from the various aquifers since major pumping began about 1900. By 1980, a total of about 204 ft^3/s was being withdrawn by major users for municipal and industrial supplies. This total was the amount simulated and represents about 60 percent of the actual pumpage. The simulated heavy pumpages used in 1980 were from the Patuxent (55 ft^3/s), Patapsco (51 ft^3/s), and the surficial (39 ft^3/s) aquifers. Major pumping centers are located south of Baltimore, Md., south and east of Washington, D.C., near Dover, Del., in southern Maryland, along both the eastern and western coast of the Chesapeake Bay, in northern Delaware, on the lower Eastern Shore of Maryland, and along the Atlantic Coast of both Maryland and Delaware. Pumpage in each of these areas is from one or more of the 10 major aquifers, and as a result, large regional cones of depression have developed. The largest and deepest simulated cones occur in the Patuxent aquifer in northern Delaware (110 ft deep) and in the Magothy aquifer in southern Maryland (50 ft deep). Potentiometric heads in these cones of depression have been lowered well below sea level.

A quasi-three-dimensional, finite-difference groundwater flow model was used to simulate ground-water flow in and between the aquifers. A 10-layer model was constructed so that each layer (except the top layer) contained a major aquifer and its overlying confining unit. The top layer, the surficial aquifer, did not include an overlying confining unit. Each layer was divided into square grid cells 3.5 mi on a side, using a 42-by-36 grid. Model inputs included the distribution and magnitude of aquifer transmissivity and storage coefficients, the distribution and magnitude of confining-unit leakance, and pumpage from each aquifer. The underlying crystalline basement rocks were considered as a no-flow boundary, as were the limits of the freshwater boundary (a line representing the 10,000-mg/L isochlor) in the southern and eastern part of the deeper aquifers and the edges of the aquifer-outcrop areas in the north and west. The northeastern and southwestern boundary conditions were modeled as constant flux, with the amounts of flux based on simulations made with the regional model of the entire northern Atlantic Coastal Plain aquifer system. The water table in the surficial aquifer and in the outcrop areas of the other aquifers was treated as a constant-head boundary during simulations of prepumping conditions.

The model was calibrated under steady-state prepumping conditions (1900) and transient pumping conditions (1900–80). In both calibration simulations, transmissivity and leakance were the only model inputs that were adjusted. Initial transmissivities were adjusted by a maximum of about 0.5 orders of magnitude, and the initial leakances were adjusted by about 3 orders of magnitude. For the transient pumping calibrations, an extra layer of cells was added above the water-table cells to simulate shallow ground-water discharge to surface-water bodies. Thus, the water table that was held constant in the steady-state simulations became an active surface in the transient simulations, and a uniform rate of recharge (15 in/yr) from precipitation was added to the model. Calibration was evaluated for each aquifer by comparing simulated prepumping and 1980 potentiometric-surface maps with prepumping and potentiometric-surface maps made from observed heads, and by comparing hydrographs based on simulated heads with hydrographs of measured heads for 121 observation wells. A successful comparison was based on the criterion that simulated heads were within 5 percent of the total head range for the entire flow system (18 ft). The general features of the ground-water flow system were reasonably well approximated during the calibration simulations.

For the prepumping simulation, the principal source (96.4 percent) of ground water was recharge from precipitation (11,419 ft^3/s). The principal discharge (96.3 percent) was to the surface-water bodies (11,392 ft^3/s).

Recharge from precipitation to individual aquifers for the prepumping simulation ranged from zero for the Piney Point–Nanjemoy aquifer (which does not crop out) to 8,384 ft^3/s for the surficial aquifer. Discharge of ground water from the aquifers to surface water ranged from zero for the Piney Point–Nanjemoy aquifer to 8,352 ft^3/s for the surficial aquifer.

The velocity and direction of ground-water flow differ among aquifers. The maximum average ground-water flow velocity for the prepumping simulation, assuming a porosity of 0.2, was in the surficial aquifer (0.19 ft/d), and the direction of flow for the surficial aquifer was locally controlled with short flow paths to nearby streams, lakes, bays, or the Atlantic Ocean. The minimum average ground-water flow velocity was about 0.01 ft/d for both the Matawan and Severn aquifers. Flow directions in both of these aquifers were partly toward the outcrop areas located in the Delaware Bay and also for the Severn aquifer in the Chesapeake Bay. The maximum ground-water flow velocities for any individual cell were in the Patapsco aquifer (5.3 ft/d) and the Patuxent aquifer (4.0 ft/d).

The ground-water budget for the transient simulation for 1978–80 suggests that the principal source of water was areal recharge (11,419 ft^3/s) and the principal sink was discharge to surface water (11,208 ft^3/s). (Evapotranspiration was not simulated.) Pumpage for the transient simulation for 1978–80 was about 204 ft^3/s. This pumpage resulted in a concomitant reduction, from the prepumping simulation, of ground-water discharge to surface water of about 184 ft^3/s and flow through the confining units of about 15 ft^3/s, or about 2 and 4 percent, respectively, of the prepumping rates.

Changes have occurred in most of the modeled aquifers, from the prepumping simulation to the 1978–80 period of the transient simulation, in the direction and velocity of ground-water flow and in the magnitude of discharge to ground-water sinks. The Patuxent aquifer for the 1978–80 period had pumpage of about 55 ft^3/s, which resulted in reversals of flow directions, especially east of Baltimore (where the original easterly direction of flow for the prepumping simulation was westerly for the 1978–80 period) and in northern Delaware. Simulated discharge to the surface-water bodies was reduced (39 ft^3/s), and average ground-water flow velocities increased (58 percent).

Simulated pumpage of 51 ft^3/s from the Patapsco aquifer reduced discharge to surface water by 46 ft^3/s. The average ground-water flow velocity remained unchanged except in northern Delaware, where the ground-water flow velocities increased about threefold. A regional change in the direction of ground-water flow toward the south resulted from large withdrawals in Virginia.

Simulated pumpage withdrawal from the Magothy aquifer for the 1978–80 period was about 14 ft^3/s. The principal sink was discharge to surface water, and this discharge decreased by about 4 percent from the prepumping to the 1978–80 period. Average ground-water flow velocities increased from 0.05 to 0.07 ft/d. The ground-water flow patterns readjusted to the location of pumpage so that much of the flow both on the Eastern Shore and in southern Maryland is toward two major pumping centers on the eastern side of the Chesapeake Bay.

Simulated pumpage in the Aquia-Rancocas aquifer (9 ft^3/s) for the 1978–80 period was widely distributed (77 model cells). The principal ground-water sink is discharge to surface water, which decreased by 6 ft^3/s from prepumping conditions. For the prepumping period, ground-water flow was to the Delaware Bay, the Chesapeake Bay, and the Potomac River. For the 1978–80 period, this pattern was greatly altered so that flow under the Chesapeake Bay and the Potomac River was toward pumping centers located in southern Maryland and on the Eastern Shore of Maryland.

The Piney Point–Nanjemoy aquifer does not crop out or subcrop in the study area. Thus, the major change from the prepumping conditions to the 1978–80 period due to simulated pumpage (10 ft^3/s) was a 9 ft^3/s net increase of flow through the adjacent confining units into the aquifer. Average simulated ground-water flow velocities increased by about 50 percent. The pattern of ground-water flow for the prepumping period was to the Chesapeake and Delaware Bays and major rivers, where the ground water discharged through overlying confining units. For the 1978–80 period, the flow pattern indicates that flow is passing under the bays to three major pumping centers.

Most of the simulated ground water discharged from the lower Chesapeake aquifer for the 1978–80 period is to surface-water bodies (259 ft^3/s). Pumpage (10 ft^3/s) has reduced this discharge by only 2 ft^3/s, but net leakage through adjacent confining units into the aquifer has increased by 9 ft^3/s.

For the 1978–80 period, simulated pumpage in the upper Chesapeake aquifer was 14 ft^3/s. The principal ground-water sources were recharge from precipitation (209 ft^3/s) and leakage through the confining units (122 ft^3/s). Pumpage (14 ft^3/s) was compensated by a 4 ft^3/s reduction of ground-water flow to surface water and a 13 ft^3/s net increase of vertical leakage through the confining units to the aquifer. Average ground-water flow velocities (0.13 ft/d) and direction of ground-water flow remained essentially unchanged from the prepumping conditions.

Simulated pumpage of 39 ft^3/s in the surficial aquifer (which is about 32 percent of actual pumpage) for the 1978–80 period is only 0.5 percent of the simulated ground-water sinks for this aquifer. The principal sink is discharge to surface-water bodies of 8,283 ft^3/s, a reduction of 69 ft^3/s from the prepumping conditions. The net change of vertical flow out of the aquifer through the underlying confining unit was 40 ft^3/s. Both the average ground-water flow velocity (0.19 ft/d) and the direction of ground-water flow remained unchanged from the prepumping conditions to the 1978–80 conditions.

A sensitivity analysis indicated the importance of accurate pumpage data. Increasing pumpage by as little as 20 percent resulted in decreased heads of up to 15 ft. A further problem indicated by the sensitivity analyses was that a significant increase in deep percolation would result in considerably higher heads. Sensitivity of the model to the transmissivity of a cell located near pumpage indicated that a fivefold decrease in transmissivity changed the head by 165 ft. A calibrated model may be used to understand the flow system under different conditions, but the sensitivity analysis indicates the possible impacts on the model results with respect to the reliability of available data.

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